

CHAPTER -4

PHYSICAL PROPERTIES OF SEA WATER

4.0: INTRODUCTION:

Measuring properties of water at various depths in the ocean helps in understanding how ocean water moves at different depths. This can be done by lowering instruments from the sea surface to sea-bottom, which continuously measure the temperature, salinity and other properties of seawater.

Water is one of the most remarkable compounds in nature. Most of the processes of mankind's environment ultimately depend on its unique physical properties. The liquid state of water is rare in the universe.

The waters of the land originate in the sea where the ocean basins hold some 1372 million cubic kilometers of salt water. From this vast store of ocean, only 3,34,000 cubic kilometers of fresh water are drawn up each year by evaporation and through the atmospheric cycle are given back to the seas by direct precipitation and through the flow of rivers. And more than 99000 km³ of rain descend each year over the continents.

Compared to other chemically related compounds water behaves physically in a unique manner. For example its freezing and boiling points should be much lower (-150 °C and -80°C) as is existing now in nature (anomalous behavior). Actually water could exist only in gaseous form. Also nearly all material substances expand when heated and contract when cooled, but water follows this rule only partly. This kind of behavior is called anomalous behavior. At temperatures below 4°C it expands with further cooling freezes at nearly 0°C, it expands suddenly by about 9%. If this abnormality were not existed, ice would have sunk down to the bottom instead of floating at the surface and thus forming a kind of "protective shield" "to prevent further freezing of the water at the poles. If ice were to sink, all the polar waters would have been gradually frozen to solid.

There are other interesting features of water that affect or control life on earth. With the exception of Ammonia, the heat capacity of water is the highest of all liquids and solids in nature. This enables water to store great quantities of heat which can be released to the atmosphere at different places and times.

Water also dissolves more substances than any other liquid in the universe; this is the reason why sea water represents a repository of a great number of chemical elements. Some of these dissolved chemical compounds provide nourishment for the growth of minute organisms and phytoplankton.

4.1. BASIC PURE WATER PROPERTIES:

Sea water is a mixture of 96.5% pure water and 3.5% other material, such as salts, dissolved gases, organic substances, and undissolved particles. Its physical properties are mainly determined by the 96.5% pure water. The physical properties of pure water will therefore be discussed first.

Pure water, when compared with fluids of similar composition, displays most uncommon properties. This is the result of the particular structure of the water molecule H_2O : The hydrogen atoms carry one positive charge, the oxygen atom two negative charges, but the atomic arrangement in the water molecule is such that the charges are not neutralized. The charges would be neutralized if the angle were 180° rather than 105° (Fig.4.1).

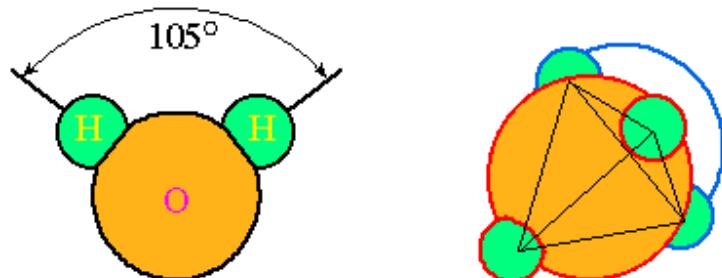


Fig.4.1. Left diagram: Arrangement of the oxygen atom (O) and the two hydrogen atoms (H) in the water molecule. The angle between the positively charged hydrogen atoms is 105° , which is very close to the angles in a regular tetrahedron ($109^\circ 28'$). Right diagram: Interaction of two water molecules in the tetrahedral arrangement of the *hydrogen bond*. The hydrogen atoms of the blue water molecule attach to the red water molecule in such a way that the four hydrogen atoms form a tetrahedron. (By courtesy of www.es.flinders.edu.au/~mattom/IntroOc/notes/figures/fig3a1.html)

The major consequences of the molecular structure of pure water are:

The water molecule is an electric dipole, forming aggregations of molecules (polymers), of on average 6 molecules at $20^\circ C$. Therefore, water reacts slower to changes than individual molecules; for example the boiling point is shifted from $-80^\circ C$ to $100^\circ C$, the freezing point from $-110^\circ C$ to $0^\circ C$.

Water has an unusually strong disassociative power, i.e. it splits dissolved material into electrically charged ions. As a consequence, dissolved material greatly increases the electrical conductivity of water. The conductivity of pure water is relatively low, but that of sea water is midway between pure water and copper. At $20^\circ C$, the resistance of sea water of 3.5% salt content over 1.3 km roughly equals that of pure water over 1 mm.

The angle 105° is close to the angle of a tetrahedron (Fig.4.2), i.e. a structure with four arms emanating from a centre at equal angles ($109^\circ 28'$). As a result, oxygen atoms in water try to have four hydrogen atoms attached to them in a tetrahedral arrangement. This is called a "hydrogen bond", in contrast to the (ionic) molecular bond and covalent bonding. Hydrogen bonds need a bonding energy 10 to 100 times smaller than molecular bonds, so water is very flexible in its reaction to changing chemical conditions.

Tetrahedrons are of a more wide-meshed nature than the molecular closest packing arrangement. They form aggregates of single, two, four and eight molecules. At high temperatures the one and two molecule aggregates dominate; as the temperature falls the larger clusters begin to dominate. The larger clusters occupy less space than the same number of molecules in smaller clusters. As a result, the density of water shows a maximum at $4^\circ C$.

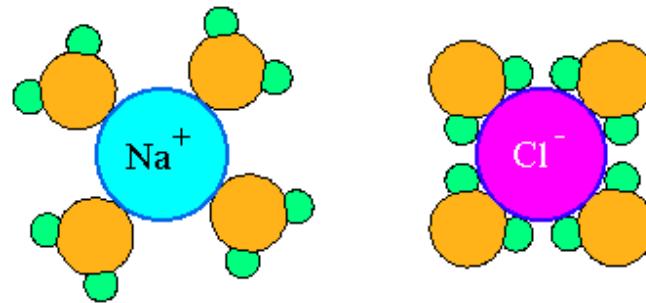


Fig.4.2. When freezing, all water molecules form tetrahedrons. This leads to a sudden expansion in volume, i.e. a decrease in density. The solid phase of water is therefore lighter than the liquid phase, which is a rare property. (By courtesy of www.es.flinders.edu.au/~mattom/IntroOc/notes/figures/.html).

Some important consequences are:

- Ice floats. This is important for life in freshwater lakes, since the ice acts as an insulator against further heat loss, preventing the water to freeze from the surface to the bottom.
- Density shows a rapid decrease as the freezing point is approached. The resulting expansion during freezing is a major cause for the weathering of rocks.
- The freezing point decreases under pressure. As a consequence, melting occurs at the base of glaciers, which facilitates glacier flow.
- Hydrogen bonds give way under pressure, i.e. ice under pressure becomes plastic. As a consequence, the inland ice of the Antarctic and the Arctic flows, shedding icebergs at the outer rims. Without this process all water would eventually end up as ice in the Polar Regions. Table 4.1 gives different properties of pure water.

The table 4.1 properties of pure water.

Molecular weight	18
Dipole moment	1.84×10^{-18}
Dielectric constant	80
Density	1.0
Boiling point	100°C
Melting point	0°C
Specific heat	1.0
Latent heat of evaporation	540 cal/g
Latent heat of fusion	79 cal/g
Surface tension at 20°C	73 dyne/cm
Viscosity at 20°C	0.01 poise
Sound velocity shows a maximum at	74°C

Do you know the sea water more easily breaks into froth or foam when disturbed than the fresh water? The reason for it is the surface tension of sea water is less than that of fresh water and so it readily breaks.

Of utmost importance to the behavior of inorganic dissolved substances is the dielectric constant. Water has the highest dielectric constant of all liquids. The dielectric constant ϵ is a number that expresses how much smaller the electric intensity is in the space filled by the dielectric than in vacuum if the same electric field is provided.

If C' is the capacitance of a condenser in vacuum, the capacitance with a dielectric between the condenser plates is given by $C = \epsilon C'$. Hence sea water acts as a good condenser with positive and negative plates as Na^{++} & Cl^- respectively (Fig.4.3).

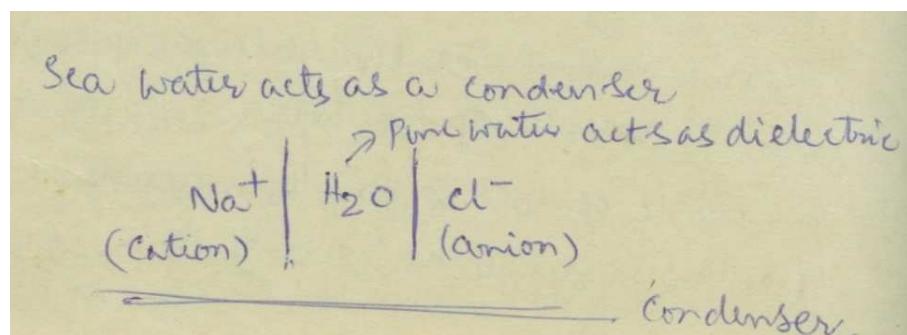


Fig.4.3. Sea water as condenser

The reason for the large value of ϵ for water lies in an abnormality of the structure of the water molecule. Since the water molecule consists of two atoms of hydrogen and one atom of oxygen, a natural possibility to build up a water molecule would be to join the two H^+ atoms to the double negative Oxygen atom (O^-) in such a way that the two valences of O^- grasp the H^+ atoms exactly at two opposite points of the O^- (180°). However, this is not the molecular structure of a water molecule. If this were the structure it's boiling and freezing points respectively would have been around -80°C and -150°C . The actual structure is that the two valences of oxygen join the valances of two hydrogen atoms at an angle of 105° to 110° . Hence the molecular structure of a water molecule is as shown in Fig.4.4

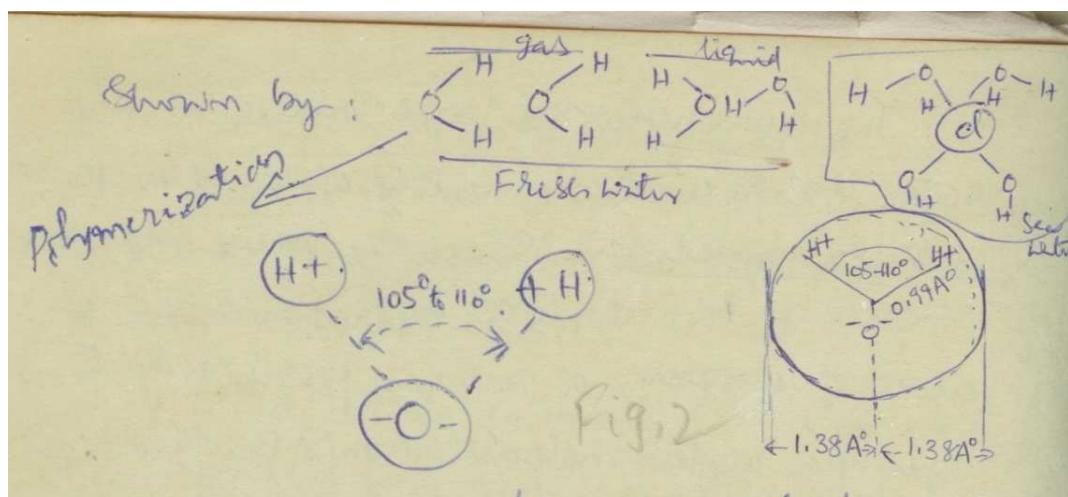


Fig.4.4. Polymerization and water molecule structure

This asymmetric structure with double negative charge results in the formation of a strong dipole moment. The dipole moment is the charge times the distance from the charge centres. This dipole moment together with the small molecular volume of the water (18) is the cause of the large dielectric constant.

The dipole moment is the cause for the existence of strong forces interacting between the molecules themselves. This is evident in the great associative power of water molecules, which leads to the formation of molecular groups like mono, di and tri hydrol groups called polymerization as shown in Fig.4.4

The considerable surface tension, viscosity, specific heat, latent heat of evaporation, high melting and boiling points etc of water as compared to other chemical compounds can be explained by the great associative power and polymerization.

There is another peculiarity of the physical – chemical behavior of water, in addition to the above, that is its anomalous behavior. That is any normal substance on heating expands and on cooling contracts. Water behaves this property partly. The reason for this is it has three different molecular structures depending on the state of water i.e solid, liquid or gas. For these three states, three different structures of tetrahedral, quartz-like lattice and ball pack structures are given.

The discovery of isotopes of hydrogen and oxygen has also modified our concept of water. We have two different masses of hydrogen (H^1 and H^2) and three different masses of oxygen (O^{16} , O^{17} , O^{18}). By combining these five different molecules; we can get nine different water molecules. Actually 99.73% of the volume of pure water consists of H_2O^{16} while all other eight possible mixtures may be present in varying proportions as H_2O^{18} , H_2O^{17} , H_2HO^{16}

The importance of anomalous of physical properties of sea water in the physical/biological environment is given in Table.4.1

Table 4.1 Importance of anomalous of physical properties of sea water in the physical/biological environment

Property	Comparison with other substances	Importance in physical/biological environment
Specific heat ($= 4.18 \times 10^3 \text{ J kg}^{-1} \text{ }^{\circ}\text{C}^{-1}$)	Highest of all solids and liquids except liquid NH_3	Prevents extreme ranges in temperature; heat transfer by water movements is very large; tends to maintain uniform body temperatures
Latent heat of fusion ($= 3.33 \times 10^5 \text{ J kg}^{-1}$)	Highest except NH_3	Absorption or release of latent heat results in large thermostatic effect at freezing point
Latent heat of evaporation ($= 2.25 \times 10^6 \text{ J kg}^{-1}$)	Highest of all substances	Absorption or release of latent heat results in large thermostatic effect at boiling point; large latent heat of evaporation is extremely important in heat and water transfer within the atmosphere
Thermal expansion	Temperature of maximum density decreases with increasing salinity; for pure water it is at 4°C	Freshwater and dilute seawater have maximum density at temperatures above the freezing point; the maximum density of normal seawater is at the freezing point
Surface tension ($= 7.2 \times 10^9 \text{ N m}^{-1}$)*	Highest of all liquids	Important in cell physiology; controls certain surface phenomena and the formation and behaviour of droplets
Dissolving power	In general, dissolves more substances and in greater quantities than any other liquid	Obvious implications in both physical and biological phenomena
Dielectric constant† ($= 87$ at 0°C , 80 at 20°C)	Pure water has the highest of all liquids except H_2O_2 and HCN	Important in the behaviour of inorganic dissolved substances because of the resulting high dissociation
Electrolytic dissociation	Very small	A neutral substance, yet contains both H^+ and OH^- ions
Transparency	Relatively great	Absorption of radiant energy is large in infrared and ultraviolet; in the visible portion of the energy spectrum there is relatively little selective absorption, hence pure water is 'colourless' in small amounts; characteristic absorption important in physical and biological phenomena
Conduction of heat	Highest of all liquids	Important on a small scale, as in living cells, but molecular processes outweighed by turbulent diffusion
Molecular viscosity ($= 10^{-3} \text{ N s m}^{-2}$)*	Less than most other liquids at comparable temperature	Flows readily to equalize pressure differences

* N = newton = unit of force in kg m s^{-2} .

† Measure of the ability to keep oppositely charged ions in solution apart from one another.

4.2: SALINITY:

Did you ever wonder why the oceans are filled with salt water instead of fresh? Just where did the salt come from? And is it the same salt you find on the dining room table? Most of the salt in the oceans came from land. Over millions of years, rain, rivers, and streams have washed over rocks containing the compound sodium chloride (NaCl), and carried it into the sea. Some of the salt in the oceans comes from undersea volcanoes and hydrothermal vents. When water evaporates from the surface of the ocean, the salt is left behind. After millions of years, the oceans have developed a noticeably salty taste.

Different bodies of water have different amounts of salt mixed in, or different salinities. Salinity is expressed by the amount of salt found in 1,000 grams of water. Therefore, if we have 1 gram of salt and 1,000 grams of water, the salinity is 1 part per thousand, or 1‰ (Parts per mille).

The average ocean salinity is 35‰. This number varies between about 32 and 37‰. Rainfall, evaporation, river runoff, and ice formation causes the variations. For example, the Black Sea is so diluted by river runoff, its average salinity is only 16‰.

Freshwater salinity is usually less than 0.5‰. Water between 0.5‰ and 17‰ is called brackish. Estuaries (where fresh river water meets salty ocean water) are examples of brackish waters.

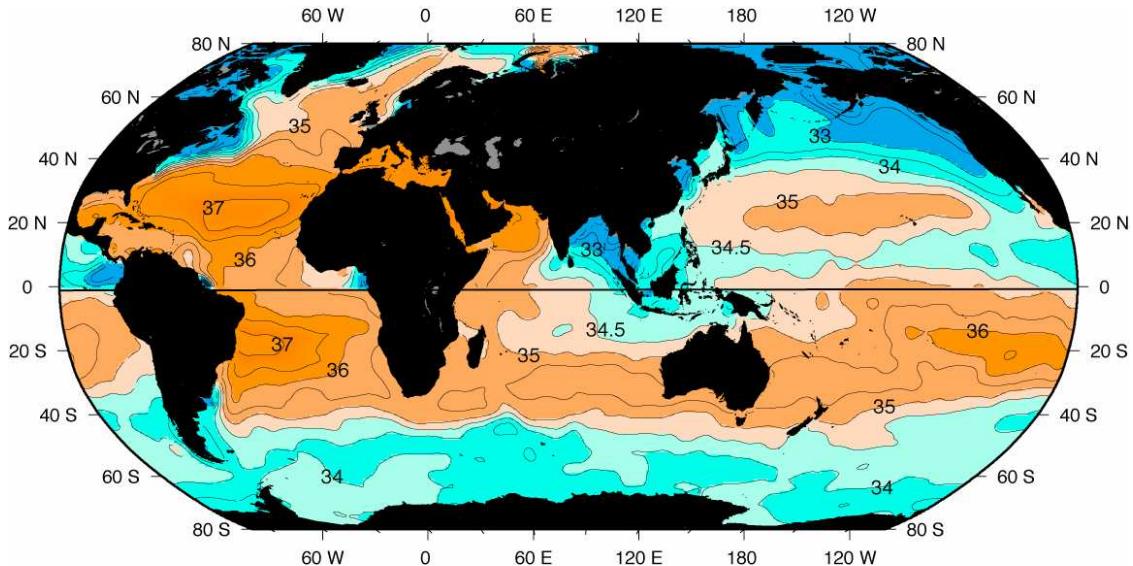


Fig 4. 5. Surface Salinity of the world oceans

The figure 4.5 is surface salinity of the world oceans. Most marine creatures keep the salinity inside their bodies at about the same concentration as the water outside their bodies because water likes a balance. If an animal that usually lives in salt water were placed in fresh water, the fresh water would flow into the animal through its skin. If a fresh water animal found itself in the salty ocean, the water inside of it would rush out. The process by which water flows through a semi-permeable membrane (a material that lets only some things pass through it) such as the animal's skin from an area of high concentration (lots of water, little salt) to an area of low concentration (little water, lots of salt) is called osmosis.

This is also why humans (and nearly all mammals) cannot drink salt water. When you take in those extra salts, your body will need to expel them as quickly as possible. Your kidneys will try to flush the salts out of your body in urine, and in the process pump out more water than you are taking in. Soon you'll be dehydrated and your cells and organs will not be able to function properly.

The sea water is a complicated solution containing 96.5% pure water and the rest 3.5% dissolved materials in the form of molecules or ions. These dissolved materials are mostly in a solution of different salts. A minor part consists of collides and suspensions. Some of these abundant ions are

Chloride ion	55% of the total dissolved material
Sodium ion	30.6% of the total dissolved material
Sulphate ion	7.7% of the total dissolved material
Magnesium ion	3.7% of the total dissolved material
Potassium ion	1.1% of the total dissolved material

The salinity can be defined as 'the total amount of dissolved material in grams contained in one kilogram of sea water when all the carbonate has been converted to oxide, the bromine and iodine are replaced by equivalent amount of chlorine and all the organic matter is completely oxidized. A significant feature of sea water is that while the total concentration of dissolved salts varies from place to place, the ratios of the more abundant components remain almost constant. For example the average concentrations of principal ions in sea water are as below:

Ion	PPM (%) by weight	
Chloride, Cl^-	18.980	
Sulphate ion, SO_4^{2-}	2.649	
Bicarbonate, HCO_3^-	0.140	
Bromide, Br^-	0.065	
Borate, H_2BO_3^-	0.026	Anions total (negative):
Fluoride, F^-	0.001	21.861%
Sodium, Na^+	10.556	
Magnesium, Mg^{++}	1.272	
Calcium, Ca^{++}	0.400	
Potassium, K^+	0.380	Cations total (positive):
Strontium, Sr^{++}	0.013	12.621%
Total salinity	34.482	34.482 ppm (%)

Please note that from the table it is clear that the proportion by weight of negative ions (21.861%) greatly exceeds that of positive ions (12.621%). But sea water is not negatively charged but it is neutral. To know the reason about the charge of sea water whether it is negatively or positively charged, we need to divide each ionic concentration by its atomic weight or molecular mass to get its ionic proportion. For example, if we do this for Na^+ ($10.6/23 = 0.46$) and Cl^- ($19/35.5 = 0.53$), their ionic proportions are almost same. When this is done for all the ions and ads + ve and -ve ions it would come equal. So it is electrically neutral.

Also note that only Ca, Na, K and Mg appear as major elements in sea water which are all derived from the common crustal rocks although they are not in the same order or in proportion. The reason for it is due to their degree of solubility and reactivity of different elements when rocks are weathered and the resulting products are carried away by rivers to sea. Many of the other commonest elements in rocks such as silicon, aluminum and iron are not soluble and so they are transported and deposited as placers in beach sands.

Thus the average salinity of sea water can be taken as 35 g/kg of sea water, usually written as 35‰

Salinity affects marine organisms because the process of osmosis transports water towards a higher concentration through cell walls. A fish with a cellular salinity of 1.8% will swell in fresh water and dehydrate in salt water. So, saltwater fish drink water copiously while excreting excess salts through their gills. Freshwater fish do the opposite by not drinking but excreting copious amounts of urine while losing little of their body salts.

Marine plants (seaweeds) and many lower organisms have no mechanism to control osmosis, which makes them very sensitive to the salinity of the water in which they live.

The main nutrients for plant growth are nitrogen (N as in nitrate NO_3^- , nitrite NO_2^- , ammonia NH_4^+), phosphorus (P as phosphate PO_4^{3-}) and potassium (K) followed by Sulfur (S), Magnesium (Mg) and Calcium (Ca). Iron (Fe) is an essential component of enzymes and is copiously available in soil, but not in sea water (0.0034ppm). This makes iron an essential nutrient for plankton growth. Plankton organisms (like diatoms) that make shells of silicon compounds furthermore need dissolved silicon salts (SiO_2) which at 3ppm can be rather limiting.

Units of salinity:

In the original definition, salinity units were ‰ (parts per thousand). This was replaced by the "Practical Salinity Unit" or PSU. Most recently, the recommendation of the SCOR working group on salinity is that salinity be unit less, as the measurement is now based on conductivity and is not precisely related to the mass of dissolved material.

Practical Salinity Unit (P.S.U):

Oceanographers had begun using conductivity meters to measure salinity (for example salinometer). The meters were very precise and relatively easy to use compared with the chemical techniques used to measure chlorinity. As a result, the Joint Panel also recommended that salinity be related to conductivity of sea water using:

$$S = -0.089\ 96 + 28.297\ 29 R_{15} + 12.808\ 32 R_{15}^2 \\ - 10.678\ 69 R_{15}^3 + 5.986\ 24 R_{15}^4 - 1.323\ 11 R_{15}^5 \\ R_{15} = C(S, 15, 0)/C(35, 15, 0)$$

where $C(S, 15, 0)$ is the conductivity of the sea-water sample at 15°C and atmospheric pressure, having a salinity S derived from the above equation, and $C(35, 15, 0)$ is the conductivity of standard "Copenhagen" sea water. Millero (1996) points out that this equation is not a new definition of salinity, it merely gives chlorinity as a function of conductivity of seawater relative to standard seawater.

Practical Salinity Scale of 1978:

By the early 1970s, accurate conductivity meters could be deployed from ships to measure conductivity at depth. The need to re-evaluate the salinity scale led the Joint Panel to recommend in 1981 (Unesco 1981; Lewis, 1980) that salinity be defined using only conductivity, breaking the link with chlorinity. All water samples with the same conductivity ratio have the same salinity. The Practical Salinity Unit of 1978 is now the official unit basing on the definition:

The definition of salinity was reviewed again when techniques to determine salinity from measurements of conductivity, temperature and pressure were developed. Since 1978, the "Practical Salinity Scale" defines salinity in terms of a conductivity ratio.

The **practical salinity**, S , of a sample of sea water, is defined in terms of the ratio R_t of the electrical conductivity of a sea water sample of 15°C and the pressure of one standard atmosphere, to that of a potassium chloride (KCl) solution, in which the mass fraction of KCl is 0.0324356, at the same temperature and pressure. The R_t value exactly equal to one corresponds, by definition, to a practical salinity equal to 35." The corresponding formula is:

$$\begin{aligned}
S &= 0.0080 - 0.1692 R_t^{1/2} + 25.3851 R_t + 14.0941 R_t^{3/2} \\
&\quad - 7.0261 R_t^2 + 2.7081 R_t^{5/2} + \Delta S \\
R_T &= C(S, t, 0)/C(KCl, t, 0) \\
\Delta S &= \left[\frac{(t - 15)}{1 + 0.0162(t - 15)} \right] + 0.0005 - 0.0056 R_t^{1/2} - 0.0066 R_t \\
&\quad - 0.0375 R_t^{3/2} + 0.636 R_t^2 - 0.0144 R_t^{5/2} \\
2 \leq S &\leq 42
\end{aligned}$$

where $C(S, t, 0)$ is the conductivity of the sea-water sample at temperature t and standard atmospheric pressure, and $C(KCl, t, 0)$ is the conductivity of the standard potassium chloride (KCl) solution at temperature t and standard atmospheric pressure. The standard KCl solution contains a mass of 32.4356 grams of KCl in a mass of 1.000000 kg of solution. An extension of the above equation gives salinity at any pressure (Millero et al 1976).

Note that in this definition, salinity is a ratio and so the symbol ($^{\circ}/_{\infty}$) is no longer used, but the earlier value of $35^{\circ}/_{\infty}$ corresponds to the same value of 35 without any units in the practical salinity scale. Some oceanographers cannot get used to write numbers without units and so for salinity they write "35 psu", where psu is meant for "practical salinity unit". As the practical salinity is a ratio and therefore does not have any units, the letters "psu" is not a unit and so meaningless. Again, minute differences occur between the old definitions and the new Practical Salinity Scale, but they are usually negligible.

The total amount of salt in the world oceans does not change except on the longest geological time scales. However, the salinity does change, in response to freshwater inputs from rain and runoff, and freshwater removal through evaporation.

Estimation of salinity:

The direct determination of salinity by evaporating sea water to dryness is too difficult to carry out as a routine process. So the method used to determine salinity from the given sea water sample is by titration with silver nitrate ($AgNO_3$) using potassium chromate or potassium dichromate as indicator.. Then scale up to the correction by the empirical relation:

$$\text{Salinity (‰)} = 0.030 + 1.80505 \times \text{chlorinity (‰)}$$

International Commission of Oceanography established this empirical relationship. Although this relationship was established well over the normal range of the concentration of sea water this is not valid for highly diluted or concentrated sea water. So it is modified in 1960 as

$$\text{Salinity} = 1.806555 \times \text{chlorinity}$$

Salinity can also be determined by other methods. Sea water is a good electrical conductor and its conductivity, which increases almost linearly with increasing salinity, can be used as a measure of salinity. This method was suggested by knudsen. But in recent years with the development of reliable and precise conductivity bridges, this has become popular and replaced the titration method. The development of salinometer is an advance in this process. But in recent years another invention is CTD (Conductivity Temperature Depth Recorder) and STD (Salinity Temperature Depth Recorder). The advantage of CTD or STD is it directly records the conductivity and temperature with depth as the probe goes into the ocean on the deck of the computer.

There are some more methods to determine salinity using chlorinity and refractive index. From the figure (4.6), if refractive index (μ) is known chlorinity can be obtained as they are linearly related. Knudsen earlier used Pycnometer for the determination of density and then later determined salinity at a given temperature of the sample. But this method is time consuming and cumbersome, though it is accurate. With the help of interferometer the refractive index of sea water sample can be determined and later can be related to salinity. This method also has not become popular as it gives an indirect method of obtaining salinity.

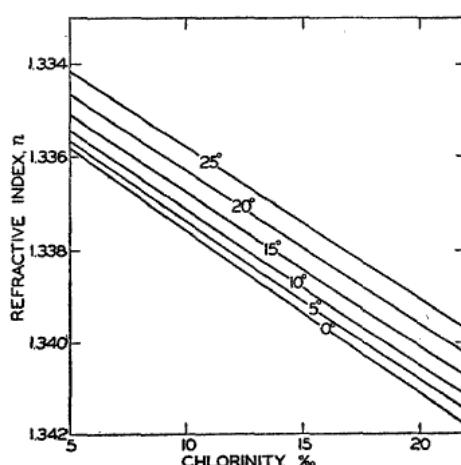


Fig.4.6. μ as a function of temperature and chlorinity

As now salinity depends on chlorinity, it is defined as the total amount of chlorine in grams present in one kilogram of sea water when bromine and iodine are replaced by an equivalent amount of chlorine.

At another term called 'chlorosity' has been in use due to the usage of word liter in place of kilogram and density of sea water is greater than unity. So chlorosity is defined as 'the total amount of chlorine in grams present in one liter of sea water when bromine and iodine are replaced by an equivalent amount of chlorine.'

$$\text{Chlorinity} = \text{Chlorosity} \div \text{density}$$

Further there is a modification in the definition of chlorinity in view of changes in the atomic weight of silver. So chlorinity once again defined as "the number giving the mass in grams of atomic weight of silver just necessary to precipitate the halogens in 0.3285233 Kgs of sea water sample".

4.2.2. Conductivity:

Conductivity of sea water depends strongly on temperature, somewhat less strongly on salinity, and very weakly on pressure (Fig.4.7). If the temperature is measured, then conductivity can be used to determine the salinity. Salinity as computed through conductivity appears to be more closely related to the actual dissolved constituents than is chlorinity, and more independent of salt composition. Therefore temperature must be measured at the same time as conductivity, to remove the temperature effect and obtain salinity. Accuracy of salinity determined from conductivity: 0.003. Precision: 0.001. The accuracy is determined by the error caused by variations in constituents such as SiO_2 which increase density but do not affect conductivity.

How is conductivity measured? (1) For a seawater sample in the laboratory, an "auto-salinometer" is used, which gives the ratio of conductivity of the seawater sample to a standard solution. The standard seawater solutions are either seawater from a particular place, or a standard KCl solution made in the laboratory. The latter provides greater accuracy and has recently become the standard. Because of the strong dependence of conductivity on temperature, the measurements must be carried out in carefully temperature-controlled conditions. (2) From an electronic instrument in the water, either inductive or capacitance cells are used, depending on the instrument manufacturer. Temperature must also be measured, from a thermistor mounted close to the conductivity sensor. Calibration procedures include matching the temperature and conductivity sensor responses.

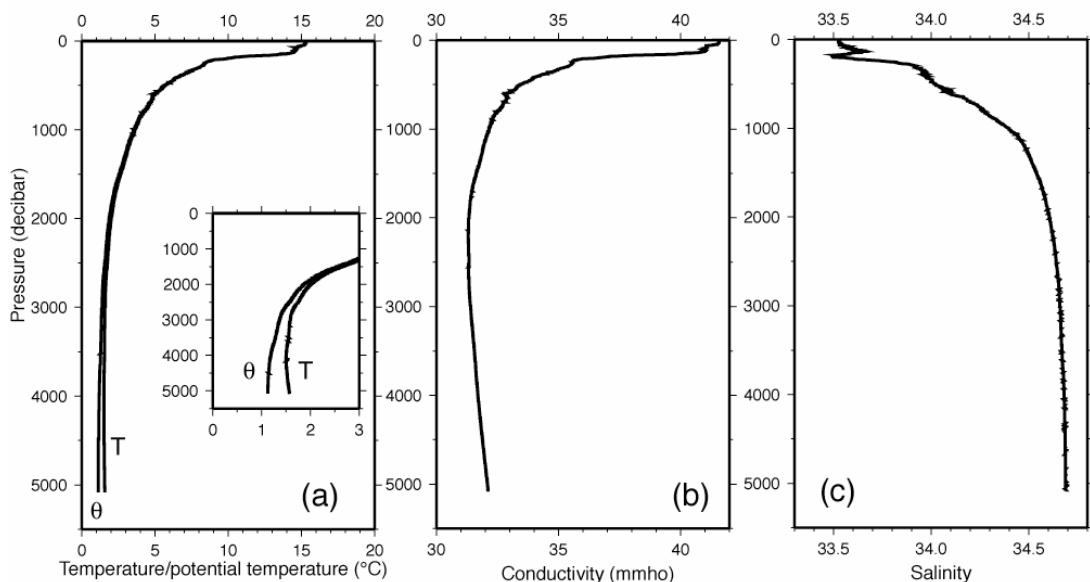


Fig.4. 7 Variation of temperature, conductivity and salinity with depth

4.3: DENSITY:

The density of any substance is defined as the mass per unit volume. Density is one of the most important parameters in the study of the ocean dynamics. Small horizontal density differences (caused for example by differences in surface heating) can produce very strong

currents. The determination of density has therefore been one of the most important tasks in oceanography. The symbol for density is the Greek letter ρ (rho).

The density of sea water depends on temperature T , salinity S and pressure p . This dependence is known as the Equation of State of Sea Water. Colder water is denser. Saltier water is denser. High pressure increases density. The dependence is nonlinear. An empirical equation of state is used, based on very careful laboratory measurements.

The equation of state for an ideal gas is given by

$$p = \rho R T$$

where R is the gas constant. Seawater is not an ideal gas, but over small temperature ranges it comes very close to one. The exact equation for the entire range of temperatures, salinities and pressures encountered in the ocean

$$\rho = \rho(T, S, P)$$

(where S is salinity) is the result of many careful laboratory determinations. The first fundamental determination to establish the equation was made in 1901 by Knudsen. His equation expressed ρ in g cm^{-3} . New fundamental determinations, based on data over a large pressure and salinity range, resulted in a new density equation, known as the "International Equation of State 1980" (Unesco 1987). This equation uses temperature in $^{\circ}\text{C}$, salinity from the Practical Salinity Scale and pressure in dbar (1 dbar = 10,000 pascal = $10,000 \text{ N m}^{-2}$) and gives density in kg m^{-3} . Thus, a density of 1.025 g cm^{-3} in the old formula corresponds to a density of 1025 kg m^{-3} in the International Equation of State.

Density increases with an increase in salinity and a decrease in temperature, except at temperatures below the density maximum. Oceanic density is usually close to 1025 kg m^{-3} (In freshwater it is close to 1000 kg m^{-3}). Oceanographers usually use the symbol σ_t (the Greek letter sigma with a subscript t) for density, which they pronounce "sigma-t". This quantity is defined as $\sigma_t = \rho - 1000$ and does not usually carry units (it should carry the same units as ρ). A typical seawater density is thus $\sigma_t = 25$.

The specific gravity on the other hand is defined as the ratio of density to that of the distilled water at a given temperature and under atmospheric pressure. In C.G.S system the density of distilled water at 4°C is equal to unity. But in oceanography specific gravity is always referred to the density of distilled water at 4°C and is therefore numerically identical to density. The reason for this is the specific weight is given instead of the density. The specific weight is defined as the quotient of the two densities (ρ / ρ_w) the substance (ρ) and distilled water (ρ_w) at a fixed temperature. The specific weight is thus a dimensionless quantity. So in C.G.S system the density and the specific gravity are numerically equal if distilled water at 4°C is taken as the comparison.

As sea water is saline, it is heavier than pure water. The density of sea water varies with salinity(s), temperature (t), and pressure (p)(Fig.4.9). So the density of sea water sample at salinity(s) temperature (t), and pressure (p) at which it is collected is called density *in situ*.

For example in the open ocean density values vary from about 1.02100 and 1.02750. For oceanographic studies it is necessary to know the density value correctly to at least five decimal

places. As it is cumbersome to carry on all the digits usually it is expressed in a compact manner highlighting the last four digits as 21.00 to 27.50 which comes from the equation:

$$\sigma = (\rho - 1) 10^3$$

For many applications in oceanography, the pressure effect on density can be ignored as the density is used for comparison at the same depth (pressure). This is known by a quantity $\sigma_{s,t,o}$ abbreviated as σ_t which is a function of salinity and temperature. Thus sigma-t is the density difference of the water sample when the total pressure on it has been reduced to atmospheric pressure while the salinity and temperature are in situ. However, for different depths we can compute σ as:

$$\begin{aligned}\sigma &= \rho(S, T, P) - 1000 \\ \sigma_t &= \rho(S, T, 0) - 1000 \quad \text{SIGMA-}t \\ &\text{Upper 1km} \\ \sigma_2 &= \rho(S, \theta, 2000) - 1000 \quad \text{SIGMA-}2 \\ &\text{1km - 3km} \\ \sigma_4 &= \rho(S, \theta, 4000) - 1000 \quad \text{SIGMA-}4 \\ &> 3km\end{aligned}$$

It is difficult to find a direct relation between σ_t salinity and temperature, as it is a complicated nonlinear function and no simple formula has been framed for it. However, it is related to another function called Sigma-0 (σ_0) which is a function of salinity alone if it is considered at, $t=0$ and on the sea surface ($p=0$). Then the empirical relation given as

$$\sigma_t = -0.093 + 0.8149 S - 0.000482 S^2 + 0.0000068 S^3$$

This also can be related to chlorinity (Fig.4.8) as

$$\sigma_t = -0.069 + 1.4708 Cl - 0.001570 Cl^2 + 0.0000398 Cl^3$$

Later σ_t and σ_0 are related by another function, D which again is a function of thermal expansion of sea water.

$$\sigma_t = \sigma_0 - D$$

All the values of σ_t , σ_0 , D are given in tables prepared by several early oceanographers such as Bjerknes and Sandström (1910), Hesselberg & Sverdrup (1914) La Fond (1951).

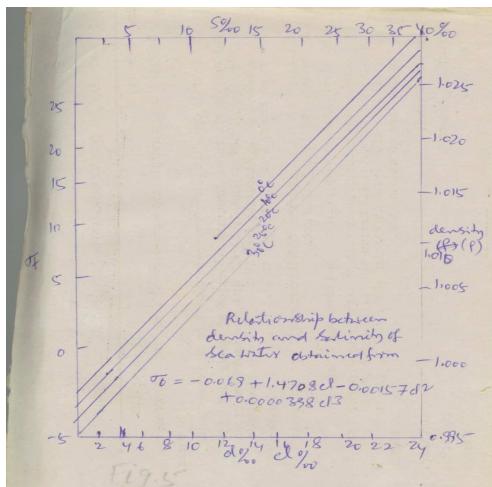


Fig.4.8 Relation between σ_t and chlorinity

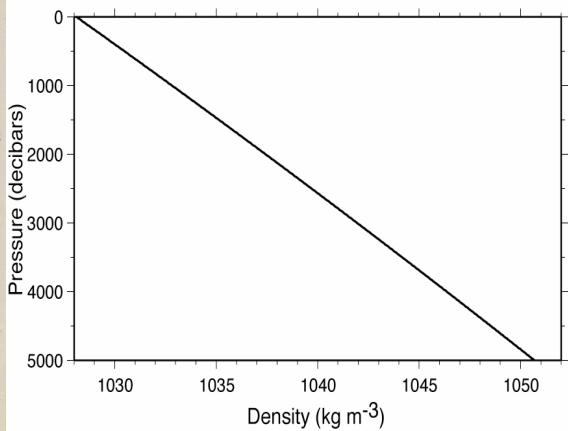


Fig.4.9 Relation between σ_t and pressure

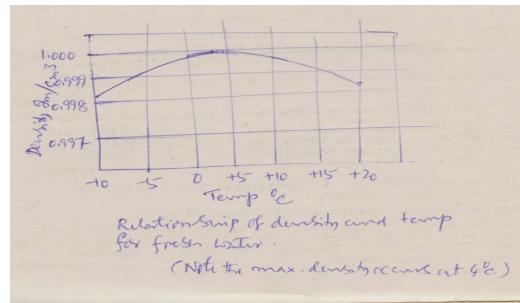


Fig.4.10 Relation between temperature and density

Fresh water ($S=0$) at atmospheric pressure ($p=0$) has maximum density at temperature 4°C . (Fig.4.10). Thus colder fresh water is less dense, which has implications for Overturn and ice floating. As salinity is increased, the density maximum moves to lower temperature. At a salinity of about 24.7‰ , the maximum density is at -1.33°C which is the freezing point of that water as shown in Fig.4.11. One may note from the Figure (4.11), 90% of the ocean water is in the ranges of $34\text{-}35\text{‰}$ salinity and $\sigma_t = 26$ to 27 and temperature -2 to 10°C . The red and blue spots denote the mean and most abundant values of temperature and salinity of the 90% of the oceans.

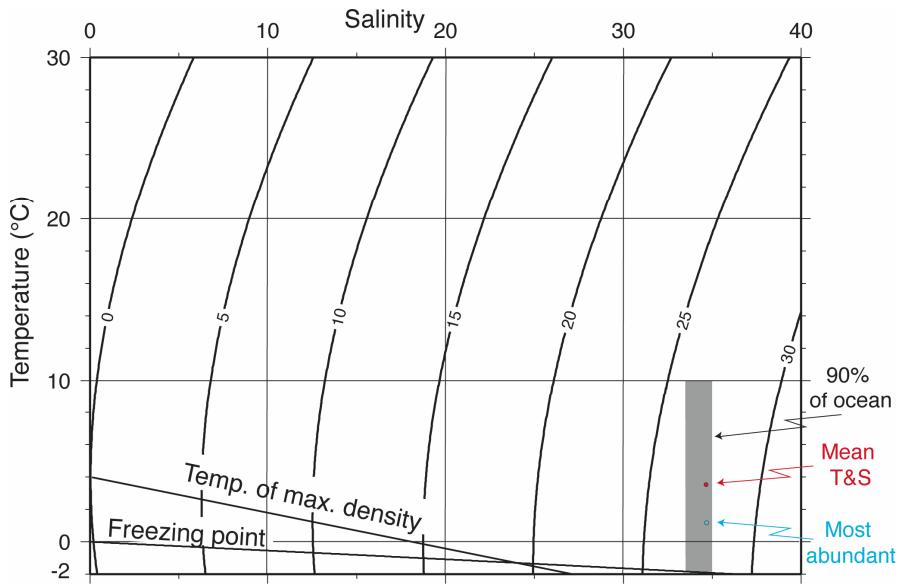


Fig.4.11 Variation of temperature of maximum density and freezing point with increase of salinity

Most variation in seawater is caused by pressure variation. This has little to do with the source of water, and if we wish to trace a water parcel from one place to another, one depth to another, we prefer to remove the pressure dependence. (This is in analogy with potential temperature).

We define potential density as the density that a parcel has when moved adiabatically to a reference pressure. If the reference pressure is the sea surface, then we compute the potential temperature of the parcel, and evaluate the density at pressure 0 dbar. But, cold water is more compressible than warm water. That is, it is easier to deform a cold parcel than a warm parcel. Therefore cold water becomes denser than warm water when they are both submerged to the same pressure. Therefore it is necessary to use a reference pressure which is relatively close to the depth we are interested in studying. The compressibility effect is apparent when we look at contours of density at say 4000 dbar compared with those at 0 dbar.

We refer to potential density at the sea surface as "sigma sub theta" σ_θ , if potential temperature is used. We refer to potential density referenced to 1000 dbar as "sigma sub 1" (σ_1), to 2000 dbar as "sigma sub 2" (σ_2), to 3000 dbar as "sigma sub 3" (σ_3) and so on, following the nomenclature introduced by Lynn and Reid (1973). In these cases, potential temperature relative to the reference pressure is used in evaluating the potential density.

The dependence of compressibility on temperature can be important. For instance, water spilling out of the Mediterranean through the Straits of Gibraltar (Fig.4.12) is extremely salty and rather warm, compared with water spilling into the Atlantic from the Greenland Sea over the Greenland-Iceland ridge. They both have about the same density at their sills. However, the warm, saline Mediterranean water does not compress as much as the Greenland Sea water, and so does not reach the ocean bottom. There is also a difference in how the two types of water entrain other waters as they plunge downwards, so this is not a straightforward explanation.

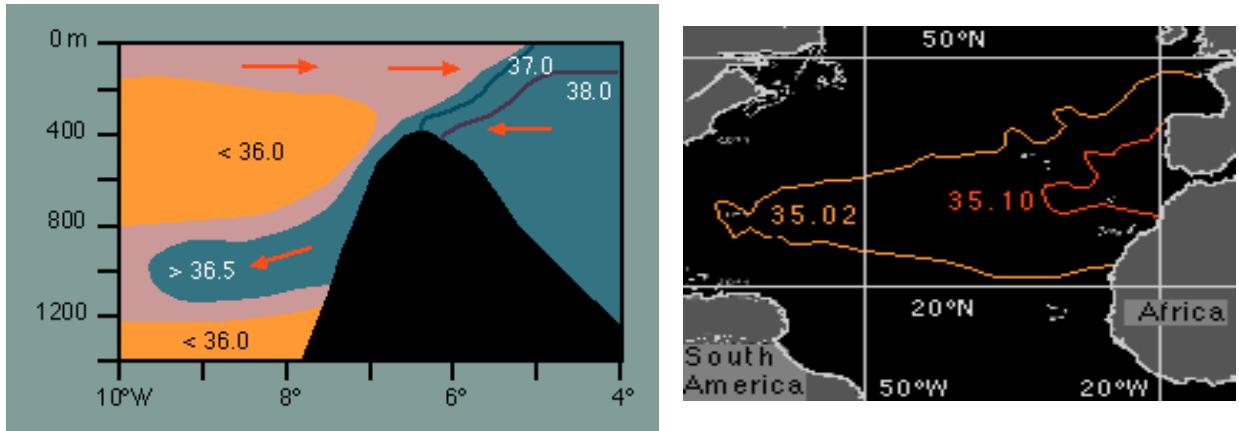


Fig.4.12 Spreading of Mediterranean Water into the Atlantic Ocean. **Left:** On entering the Atlantic Ocean across the sill, the salty (dense) Mediterranean water sinks to a depth of about 1000 m, where its density matches the density of the surrounding waters. **Right:** Areal spreading of Mediterranean water into Atlantic and the associated surface salinity distribution in Atlantic off Gibraltar.

Neutral density.

When analyzing properties in the ocean to determine where water parcels originated, it is assumed that most motion occurs with very little change in the density of the parcel, with the exception of changes due to pressure. This concept is essentially a statement that water follows an isentropic surface if it moves with no exchange of heat and salt. Defining an isentropic surface in the presence of mixing presents some difficulties. The isopycnal surfaces which we use in practice to map and trace water parcels should approximate isentropic surfaces. We typically use a reference pressure for the density which is within about 500 meters of the pressure of interest. (This pressure interval has just been found through experience to be adequate.) Therefore when working in the top 500 meters, we use a surface reference pressure. When working at 500 to 1500 meters, we use a reference pressure of 1000 dbar, etc. This discretization takes care of most of the problems associated with the effect of pressure on density. When isopycnals cover a greater range of pressure, then they must be patched into the shallower or deeper range - this is the practice followed by Reid in his various monographs on Pacific and Atlantic circulation.

A more continuously varying reference pressure for isopycnal surfaces was introduced and is referred to as a "neutral surface". If a parcel is followed along its path, assuming the path is known, and then it is possible to track its pressure continuously and continuously adjust its reference pressure and density.

The neutral surface concept was refined and a computer program is available now for computing neutral density. Neutral density depends on location in latitude/longitude/depth and is based on marching outwards around the world from a single point in the middle of the ocean, using a climatological temperature/salinity data set, and tracking imaginary parcels along radiating lines. Initial usage indicates that neutral density as determined from this program can successfully replace the approximate neutral surfaces produced by adjusting reference pressures at every 1000 dbar.

Shear flow adjustment (dynamic stability) occurs when vertical mixing in the ocean is dominated by the influence of the density gradients that limit vertical motions. The dynamic

effect of density gradient is contained in the parameter, $N = \left(-\frac{g}{\rho_0} \frac{\partial \rho}{\partial z} \right)^{1/2}$ the Brunt-Väisälä (buoyancy) frequency, which is the frequency with which a displaced element of fluid will oscillate, where ρ and ρ_0 are the densities at the observed level and surface. The corresponding periods $\frac{2\pi}{N}$ are typically a few minutes in the thermocline and up to many hours in the weakly stratified deep ocean.

4.4. SPECIFIC VOLUME:

But very often the reciprocal of density ($1/\rho$), the specific volume (α) is used for various purposes in oceanography than density. As this is also a function of salinity, temperature and pressure and required correct up to fifth decimal place, the specific volume is expressed as an anomaly (δ) and is found as the difference between *insitu* and standard as:

$$\delta = \alpha_{s,t,p} - \alpha_{35,0,p}$$

Where $\alpha_{s,t,p}$ is specific volume in situ and $\alpha_{35,0,p}$ is specific volume of an arbitrary standard ocean which has $S = 35\%$, $t = 0$ and $p = p$. In other words

$$\alpha_{s,t,p} = \alpha_{(35+\Delta s), (0+\Delta t), p}$$

Where $\alpha_{35,0,p}$ is a constant and has a value equal to 0.97264. Bjerknes and Stand Ström (1910) examined the experimental values of δ as functions of salinity, temperature and pressure and showed that δ could be broken down into number of individual components that depends on one or two parameters as:

$$\delta = \delta_s + \delta_t + \delta_{s,t} + \delta_{s,p} + \delta_{t,p}$$

where the suffixes denote the functions of the denoted parameters only. Montgomery and Wooster (1954) called the first highlighted three terms on the right of the equation as thermosteric anomaly,

$$\Delta_{s,t} = \delta_s + \delta_t + \delta_{s,t}$$

The thermosteric anomaly ($\Delta_{s,t}$) is defined as the anomaly of specific volume that would be attained if the water were changed isothermally to a standard pressure of one atmosphere.

Montgomery and Wooster (1954) further pointed out that in actual practice the effect of the sum of these three terms (the thermosteric anomaly) is adequate in most practical cases to describe the specific volume of water masses in the upper layers of the oceans as the rest of the terms ($\delta_{s,p}, \delta_{t,p}$) are negligibly small.

The relation between $\Delta_{s,t}$ and σ_t :

We know

$$\begin{aligned}
 \sigma_{s,t,o} &= \sigma_t \\
 \text{or } \sigma_{s,t,o} &= (\rho_{s,t,o} - 1)10^3 \\
 \sigma_{s,t,o} &= \left(\frac{1}{\alpha_{s,t,o}} - 1 \right)10^3 \\
 \frac{1}{\alpha_{s,t,o}} &= 1 + (\sigma_{s,t,o})10^{-3} \\
 \alpha_{s,t,o} &= \frac{1}{1 + (\sigma_{s,t,o})10^{-3}} \\
 \alpha_{s,t,o} &= 1 - \frac{10^{-3} \sigma_{s,t,o}}{1 + 10^{-3} \sigma_{s,t,o}} = 1 - \frac{10^{-3} \sigma_t}{1 + 10^{-3} \sigma_t}
 \end{aligned}$$

We further know

$$\begin{aligned}
 \alpha_{s,t,o} &= \alpha_{35,0,p} + \Delta_{s,t} = 0.97264 + \Delta_{s,t} \\
 \Delta_{s,t} &= \alpha_{s,t,o} - 0.97264 \\
 \Delta_{s,t} &= 1 - \frac{10^{-3} \sigma_t}{1 + 10^{-3} \sigma_t} - 0.97264 \\
 \Delta_{s,t} &= 0.02736 - \frac{10^{-3} \sigma_t}{1 + 10^{-3} \sigma_t}
 \end{aligned}$$

This is the relation between thermosteric anomaly and sigma-t.

Although units of α , δ and $\Delta_{s,t}$ are in Cm^3/g , $\Delta_{s,t}$ is usually represented in centiliters per ton (Cl/t). Their orders of magnitude in the oceans are as below:

$$\Delta_{s,t} = 50 \text{ to } 100 \times 10^{-5} \text{ Cm}^3/\text{g} = 50 \text{ to } 100 \text{ Cl/t}$$

$$\delta_{s,p}, \delta_{t,p} = 5 \text{ to } 15 \times 10^{-5} \text{ Cm}^3/\text{g} = 5 \text{ to } 15 \text{ Cl/t}$$

Further $\Delta_{s,t}$ and σ_t are related as: $\Delta_{s,t} = 0.02736 - (10^{-3} \sigma_t) / (1 + 10^{-3} \sigma_t)$

4. 5. THERMAL EXPANSION:

The volume of a substance generally increases with increase of temperature are called 'cubical expansion'. The relative change of volume with temperature is denoted by a term called coefficient of thermal expansion (β)

$$\beta = \frac{1}{\alpha_{s,t,p}} \frac{d\alpha_{s,t,p}}{dt} \dots \dots \dots (4)$$

Where $\alpha_{s,t,p}$ is the specific volume of sea water. The β values of sea water are greater than that of pure water and increases with increase of temperature. Negative values of β indicates that the liquid contracts with increase of temperature between 0 and 4°C.

4. 5.1. COEFFICIENT OF THERMAL CONDUCTIVITY:

The amount of heat in gram calories per second which is conducted through a surface area of 1 cm^2 is proportional to the change in temperature per centimeter along a line normal to that surface and the coefficient of proportionality ' γ ' is called the coefficient of thermal conductivity.

$$\frac{dq}{dt} \propto A \frac{dT}{dn} = -\gamma A \frac{dT}{dn}$$

$$\therefore \gamma = -\frac{1}{A} \frac{dq/dt}{dT/dn} \text{ cal cm}^{-2} \text{ sec}^{-1} / \text{oC} \quad \dots\dots\dots (5)$$

Where dq = amount of heat in Calories, dt = change in time, dT = change in temperature, dn = normal to the surface and A = exchange coefficient. The negative sign indicates that the heat flux is considered positive if it is in the direction of decreasing temperature along the normal 'n'. The proportionality constant 'γ' is called the coefficient of thermal conductivity and is expressed as $\text{cal/cm}^2/\text{sec}^0\text{c}$. γ value for pure water is $1.39 \times 10^3 \text{ cal/cm}^2/\text{sec}^0\text{c}$. This coefficient decreases with increase of salinity.

The variation of coefficient of thermal expansion of sea water at 35‰ increases with both temperature as well as pressure

4.5.2. SPECIFIC HEAT:

It is defined as the number of calories of heat that is required to raise the temperature of one gram of substance to one degree centigrade.

It is important to note that there are two specific heats with respect to pressure and volume viz, C_p and C_v respectively. The unit is cal/gm°c. C_p for pure water is unity. The specific heat of sea water depends on temperature, salinity and pressure.

The effect of pressure on specific heat has been given by Ekman(1914) as

$$\frac{dC_p}{dp} = -\frac{T}{\rho J} \left(\frac{d\beta}{dt} + \beta^2 \right) \dots \dots \dots (6)$$

Where $T = 273+t$ and β is the coefficient of thermal expansion. For sea water less than 20%, C_p decreases with increase of temperature.

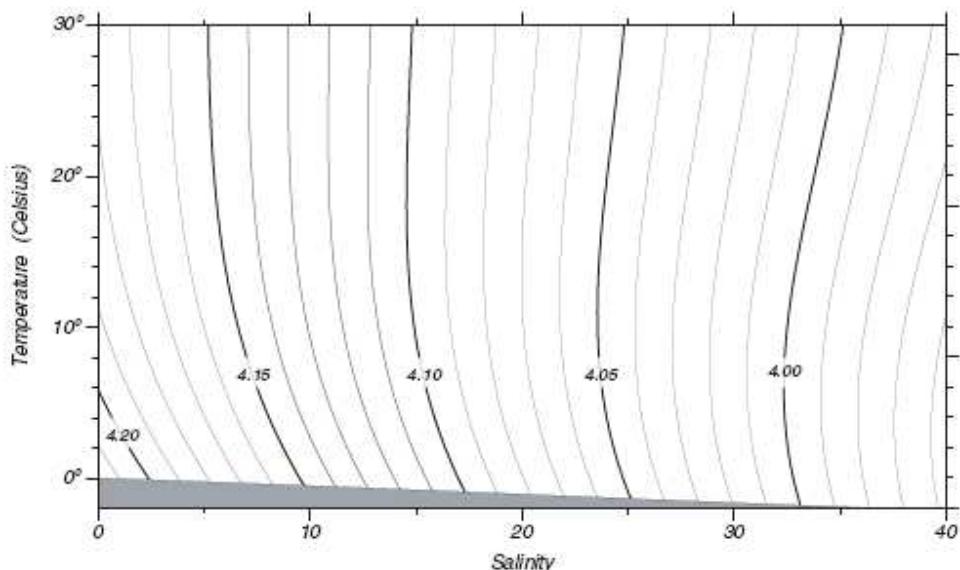


Fig.4.13. Specific heat of sea water at atmospheric pressure C_p in Joules per gram per degree Celsius as a function of temperature and salinity, calculated from the empirical formula given by Millero et al (1976) using algorithms in Fofonoff et al (1974). The lower line is freezing point of salt water.

Units for heat fluxes are watts/m². The product of flux times the surface area times time is energy in joules. The change in temperature Δt of the water is related to change in energy ΔE through:

$$\Delta E = C_p m \Delta t$$

where m is the mass of water being warmed or cooled, and C_p is the specific heat of sea water at constant pressure.

$$C_p \approx 4.0 \times 10^3 \text{ J} \cdot \text{kg}^{-1} \cdot ^\circ\text{C}^{-1}$$

Thus, 4,000 joules of energy are required to heat 1.0 kilogram of sea water by 1.0°C (figure 4.13).

4.5.3. ADIABATIC TEMPERATURE CHANGES OF SEA WATER:

If a work is done on the parcel of sea water without the gain or loss of heat to its environment – such a process is called adiabatic process. Adiabatic process means no heat is added or removed during the process.

As sea water is considered as incompressible for all practical purposes, the pressure changes undergone by a small parcel of water in the ocean must be accompanied by adiabatic changes in temperature which can be significant for oceanographic problems.

Suppose if a parcel of water is raised from certain depth adiabatically, it is cooled as it is subjected to lesser sea pressure than at its original position. As a result it expands performing work against the external pressure and so water is cooled by a certain amount. Alternatively if a water mass sinks adiabatic heating takes place.

Since the compressibility of water is not large, these temperature changes will remain only in limits. However, since the vertical temperature gradient in the deeper layers is extremely small, these adiabatic effects must be taken into account.

The adiabatic or potential temperature change $d\Theta$ for a displacement from a depth h_1 to a depth h_2 can be calculated using a formula from the energy principle derived by Lord Kelvin.

$$dQ = C_p dT - \frac{T}{J} \frac{\partial \alpha}{\partial t} dp \quad \dots \dots \dots (7)$$

Where α is specific volume of sea water and J is mechanical equivalent of heat = 4.1863×10^7 ergs/cal or dyn cm/cal.

$$\beta = \frac{1}{\alpha} \frac{\partial \alpha}{\partial t}$$

From the equation (4) of coefficient of thermal expansion, $\beta = \frac{1}{\alpha} \frac{\partial \alpha}{\partial t}$ and equation (6) of specific heat $\frac{\partial C_p}{\partial p} = -10^5 \frac{T}{\rho J} \left(\frac{\partial \beta}{\partial t} + \beta^2 \right)$ where p is pressure in decibars so conversion factor 10^5 appeared, ρ is the density of sea water, $\frac{\partial \beta}{\partial t}$ is the rate of change of thermal expansion.

Also taking the hydrostatic equation, $dp = \rho g dz$, equation (7) can be written as:

For adiabatic process, $dQ = 0$, $\alpha = 1/\rho$ and T is replaced by θ , equation (7) turns out as

$$C_p d\theta - \frac{T}{J} (\alpha \beta) (\rho g dz) = 0$$

$$d\theta = \frac{T}{C_p J} \beta g dz \quad \dots \dots \dots (8)$$

$$\theta = \int d\theta = \int_{h_1}^{h_2} \frac{T}{C_p J} \beta g dz \quad \text{for the column } h_1 \text{ to } h_2$$

Thus the adiabatic temperature change $d\theta$ (equation 8) depends on coefficient of thermal expansion β , specific heat C_p and temperature T , salinity S and pressure ($h_2 - h_1$).

4.5.4. Potential temperature

We know pressure in the ocean increases greatly downward. A parcel of water moving from one pressure to another will be compressed or expanded. When a parcel of water is compressed adiabatically, that is, without exchange of heat, its temperature increases. (This is true of any fluid or gas.) When a parcel is expanded adiabatically, its temperature decreases.

The change in temperature which occurs solely due to compression or expansion is not of interest to us as it does not represent a change in heat content of the fluid. Therefore if we wish to compare the temperature of water at one pressure with water at another pressure, we should remove this effect of adiabatic compression/expansion.

The potential temperature is the temperature of a parcel of seawater which is moved from the depth of measurement to the sea surface ($P=0$) by an adiabatic process. The potential temperature is a little lower than the measured or in-situ temperature.

In the ocean, we commonly use the sea surface as our "reference" pressure for potential temperature we compare the temperatures of parcels as if they have been moved, without mixing or diffusion, to the sea surface. Since pressure is lowest at the sea surface, potential temperature (computed at surface pressure) is always lower than the actual temperature unless the water is lying at the sea surface.

Helland – Hansen (1930) and later Cox and Smith (1959) prepared tables from the above equation giving directly adiabatic heating and cooling in sea water when raised or lowered from some depth to the surface and vice versa. With these tables any adiabatic change can easily be determined. The distribution of potential temperature at 500 meters in the oceans is given in Figure 4.14. In the Indian Ocean and Bay of Bengal it is around 10 to 12°.

Example: At a depth of 9788 meters in Philippines Trench the temperature and density are measured as $T=2.6^{\circ}\text{C}$ and $\sigma=28$. What would be its potential temperature if a parcel of water is raised to the surface? Calculate from the table 4.2 below: (Ans: 1.32°C)

Table 4.2. $S = 34.85\%$, $\sigma = 28.0$

Depth(m)	$\Theta_m (^{\circ}\text{C})$						
	-2	0	2	4	6	8	10
1000	2.6	4.4	6.2	7.8	9.5	11.0	12.4
2000	7.2	10.7	14.1	17.2	20.4	23.3	26.2
3000	13.6	18.7	23.6	28.2	32.7	37.1	41.2
4000	21.7	28.4	34.7	40.6	46.3	51.9	57.2
6000	42.8	52.2	61.1	69.4	-	-	-
8000	-	81.5	92.5	102.7	-	-	-
10000	-	115.7	128.3	140.2	-	-	-

(B) Adiabatic temperature change (in 0.01°C) for the upper 1000 m at different salinities

S‰	Θ_m (°C)					
	0	4	8	12	16	20
30	3.5	7.0	10.3	13.2	16.1	18.9
32	3.9	7.3	10.6	13.5	16.4	19.1
34	4.3	7.7	10.9	13.8	16.6	19.3
36	4.7	8.1	11.2	14.1	16.9	19.6
38	5.1	8.4	11.6	14.4	17.2	19.8

Table 4. 2: (A Top) Adiabatic cooling (0.01°C) resulting from an ascent of a water particle of temperature (Θ_m) up to the sea surface, ($\sigma_i=28.0$, $S = 34.85\text{‰}$) and (b) Adiabatic temperature change (in 0.01°C) for the upper 1000 m at different salinities.

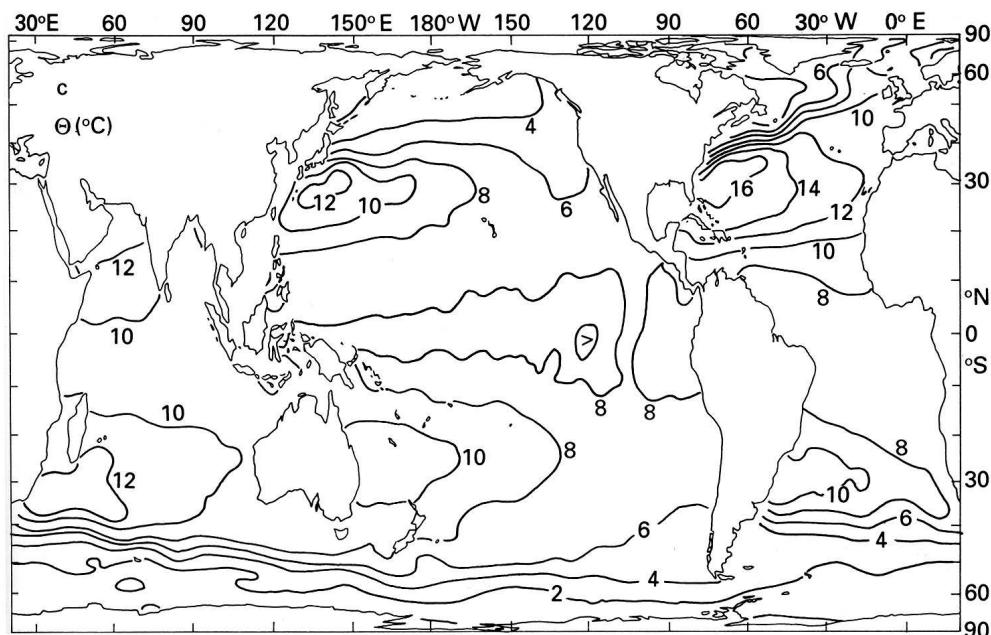


Fig.4.14 Distribution of Potential Temperature in the oceans at 500m depth

4.6. COLLIGATIVE PROPERTIES:

The colligative properties are the unique properties of solutions. Since sea water is a solution of salts and water, these properties are applicable for sea water also. The magnitude of colligative properties depends on the concentration of ions in the solution and their activity. The colligative properties are: i) Elevation of boiling point, ii) Elevation of osmotic pressure, iii) Elevation of surface tension, iv) Lowering of freezing point, v) Lowering of vapor pressure and vi) Lowering of temperature of maximum density.

4.6.1. Elevation of Boiling point: The boiling point increases linearly with increase of salinity

4.6.2. Depression of Freezing point: $T_f = -0.096 Cl - 0.0000052 Cl^2$. From this equation it is clear that freezing point, T_f decreases with increase of chlorinity (salinity). When ice is formed at polar seas, concentration of salinity increases, hence the formation of additional ice can take place only at a lower temperature.

4.6.3. Lowering of Vapor pressure: Vapor pressure of sea water of any chlorinity is referred with respect to that of distilled water as : $e/e_o = 1 - 0.000969 Cl$, where e/e_o is relative vapor pressure such that 'e' is the vapor pressure of sea water and 'e_o' is that of distilled water at the same temperature.

4.6.4. Increase of osmotic pressure: Osmotic pressure (O.P) at 0°C can be calculated from : $OP_o = -12.08 \Delta T_f$, where ΔT_f is the depression of freezing point. O.P at any temperature can be calculated: $OP_t = OP_o \left[\frac{273+t}{273} \right]$, O.P increases with increase of salinity. O.P also increases with increase of temperature.

4.6.5. Elevation of surface tension: Fleming and Revelle found the empirical relationship between surface tension, temperature and chlorinity as:

$$\Gamma \text{ (dynes/cm)} = 75.64 - 0.144 t^\circ\text{C} + 0.0399 cl\%$$

Thus it decreases with temperature and increases with salinity.

4.6.6. Lowering of temperature of maximum density:

Pure water has a maximum density at 4°C. With increasing salinity the temperature of maximum density of sea water decreases as shown by curve I of Fig.4.15. And with increasing salinity the temperature of freezing point also decreases as shown by curve II. Both the curves meet at a point (24.67, -1.33).

Thus the rate of decrease of maximum density is greater than that of freezing point upto this point (24.67, -1.33) beyond that it decreases quicker than the other one.

But at a salinity of 24.67% the temperature of maximum density and temperature of freezing point both are same (-1.33°C).

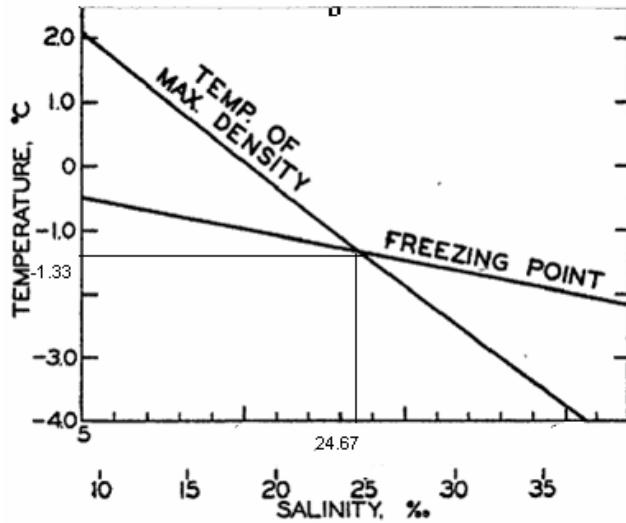


Fig.4.15 Lowering of temperature of maximum density and freezing point with increase of salinity

4.7. OPTICAL PROPERTIES OF SEA WATER.

4.7.1. Extinction of radiation:

The radiation that penetrates into the surface layers of the ocean depends upon absorption of pure water, suspended and dissolved substances, scattering by suspended and dissolved materials.

Two important terms, radiance and irradiance are usually used in this regard.

Radiance: Energy per second received by the sea surface from a certain direction.

Irradiance: Energy per second received by the sea surface from all directions that comes usually in the horizontal.

The Fig.4.16 shows the penetration of radiant energy and absorption of different wavelengths in the ocean.

Why the color of light below the sea surface is predominantly blue-green? Which wave lengths are the first to be absorbed? What proportion of the total incident energy reaches a depth of 100 m?

The Fig.4.16 gives answers to these questions. The shorter wave lengths (i.e. those near the blue end of the spectrum) penetrate deeper than longer wave lengths. So the color of light below the sea surface will be blue-green. Infrared radiation is the first to be absorbed, followed by red and so on. The total energy received at a given depth is equal to the area enclosed by the curve of that depth. The 100 m curve occupies only 1/50th of the surface curve. All of the infrared radiation is absorbed with in about a meter of the surface, and nearly half of the total incident solar energy is absorbed with in 10 cm of the surface. Penetration will also depend on the clarity or transparency of the water, which in turn depends on the amount of suspended matter in it.

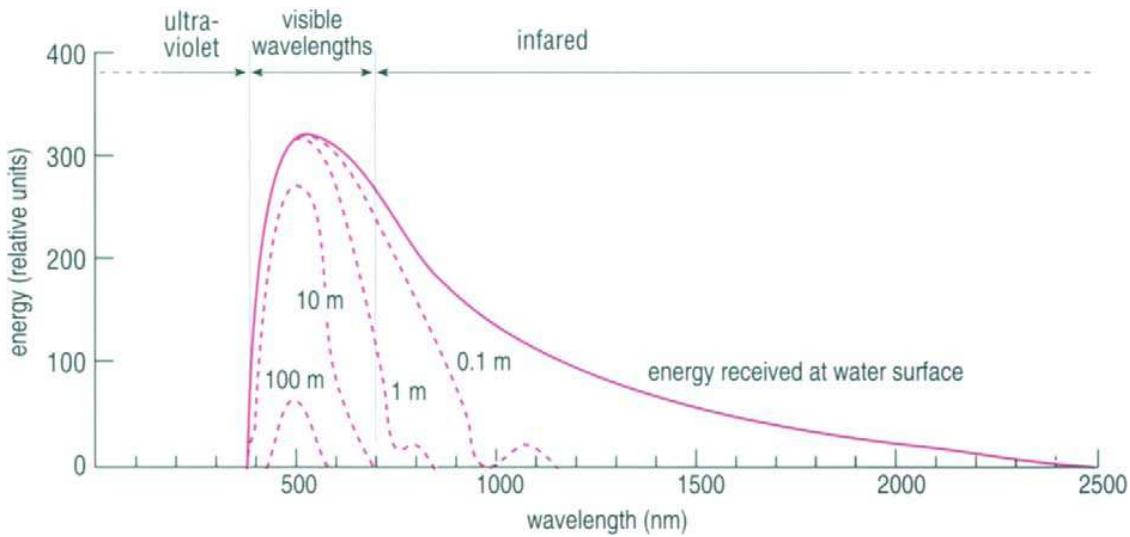


Fig.4. 16. Penetration of solar radiation at different depths in the ocean

The word used for understanding the amount of suspended and dissolved material in sea surface water is known as 'turbidity'. The turbidity is measured with the help of two instruments. One is called the 'Secchi Disc' and another is called 'Hydro photometer' or 'Transparency Gauge'. While the Secchi disc measures the vertical variation of turbidity the hydro photometer measures the horizontal variation of turbidity.

4.7.2. EXTINCTION COEFFICIENT:

The diminishing effect of radiation while penetrating into different layers of ocean is known as extinction of radiation. The extinction causes due to the attenuation of radiation which is the total effect of absorption, scattering and reflection of radiation penetrating into the layer.

In oceanography the greater interest is attached to the rate at which the downward traveling light decreases. This rate of decrease can be defined by a coefficient called extinction coefficient 'K'.

If a parallel beam of light intensity, I_0 penetrates the sea surface, the intensity I_z at a depth 'z' is given by $I_z = I_0 e^{-kz}$ Where I_0 is the amount of sun light intensity reaching the sea surface. I_z is the light intensity reaching at z meters and 'k' is the extinction coefficient.

$$-Kz = \ln(I_z) - \ln(I_0) \text{ or } K = (1/Z) 2.303 (\log I_0 - \log I_z)$$

$$\text{If } Z = 1 \text{ metre, } K = 2.303 (\log I_0 - \log I_z).$$

The factor 2.303 is due to change of natural logs to base 10 logs. Z is generally taken as one meter because it is easy for calculation and that is the reason why the two chambers of hydro photometer are kept at a distance of one meter.

The degree of light penetration varies with water type and for some water types only 5% of the light penetrates to a depth of 20 m. Fig 4.17 shows the light intensity plotted on a logarithmic scale down to 1500 m depth.

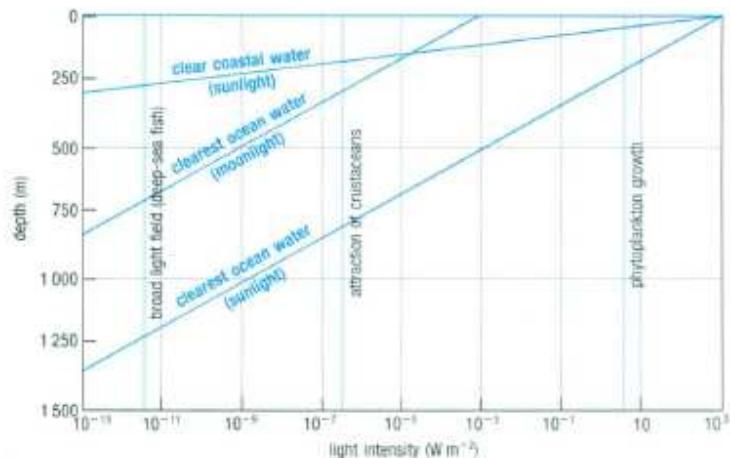


Fig. 4. 17 .Intensity of penetration of light at different depths in the ocean

The curves show the exponential decrease of the intensity of sunlight with depth, even in clearest ocean water, is such that at about 250 m depth the light intensity has fallen by nearly three orders of magnitude, from $1\ 000 (10^3) \text{ W m}^{-2}$ to little more than 1 W m^{-2} .

The pale blue vertical lines show the light intensity required for various functions. The 'broad light field' for deep-sea fish indicates the minimum quantity of general daylight that these fish can perceive. The intersection with the 'clearest ocean water (sunlight)' line at a point corresponding to about 1 250 m indicates that below this depth fish cannot perceive daylight. More light is needed to attract crustaceans, and still more for phytoplankton growth. (For comparison, the lowest intensities that the human eye can perceive are of the order of 10^{-2} W m^{-2} for a small light source and 10^{-8} to 10^{-9} W m^{-2} for a broad, diffuse light source).

4.7.3. MEASUREMENT OF EXTINCTION COEFFICIENT:

A Secchi disc (Fig.4.19) contains a brass circular disc of 30 cm radius alternatively painted black and white or some times completely white in order for visual appearance in the under water. This disc is very light and will be lowered along the plumb line vertically down ward with a suspended lead weight with the help of rope in the sea or shallow coastal bodies like Lakes, estuaries or lagoons. The depth of disappearance and appearance is noted and the average depth is called 'secchi depth, D'and this depends on the turbidity of the water. More turbid waters have less secchi depth and vice versa.

The Secchi disc is both cheap and easily made, and it has been used by oceanographers for over a century as a rapid means of assessing water quality. Simple empirical equations enable

a good deal of information to be gleaned from the Secchi depth. The basic relationship for the vertically observed Secchi disc is:

$$Z_s = \frac{F}{C + K}$$

where: Z_s is the Secchi depth; C is the attenuation coefficient for directional light; K is the extinction coefficient; and F is a factor that depends on the reflectivity of the disc and that of the background, and the observer's own threshold perception of contrast. It is about 8.7 in clear oceanic water, but can be as little as 6 in turbid estuarine water.

The empirical equation worked out for Indian waters is given in equation (9) is used to compute the extinction coefficient (K) which is directly proportional to turbidity.

$$K = 1.7/D \quad \dots \dots \dots (9)$$

The hydrophotometer or transparency gauge (Fig.4.18 & 4.21) is a heavy instrument that contains two glass chambers on either side of a one meter long carrier frame. One chamber contains a light unit (source bulb) and the other chamber contains a 'photo cell' (a detector unit). These two units are connected to a micro ammeter and the entire system works with a car battery (12V). As constant voltage is fed to the source light, after the beam of light traveling through the water column and absorbed the remaining intensity falls on the photocell which is detected by the photocell and this is measured by the micro ammeter in the deck unit. The difference in the intensity between the source light and the photo cell is proportional to the extinction coefficient and is also proportional to turbidity. A graph drawn between current and turbidity will give the variation of turbidity in different layers and columns.

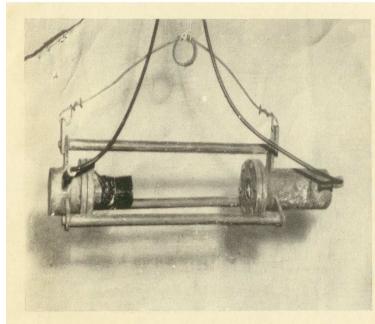


Fig.4.18 Hydrophotometer

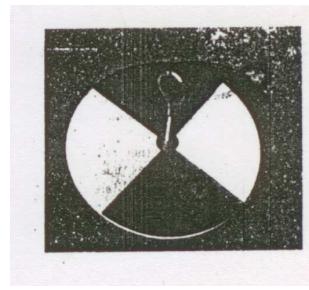


Fig.4.19 . Secchi disc

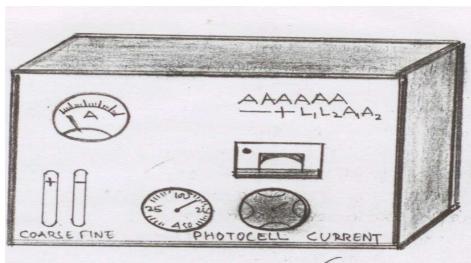


Fig. 4.20. deck unit of hydro-photometer

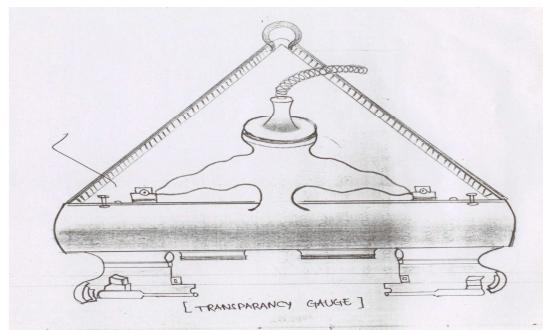


Fig. 4.21 Transparency gauge(hydrophotometer)

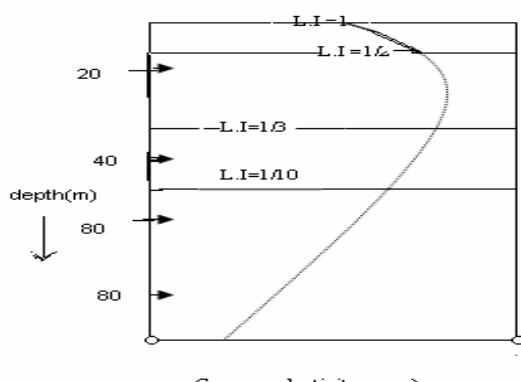


Fig. 4.22 Light intensity versus productivity

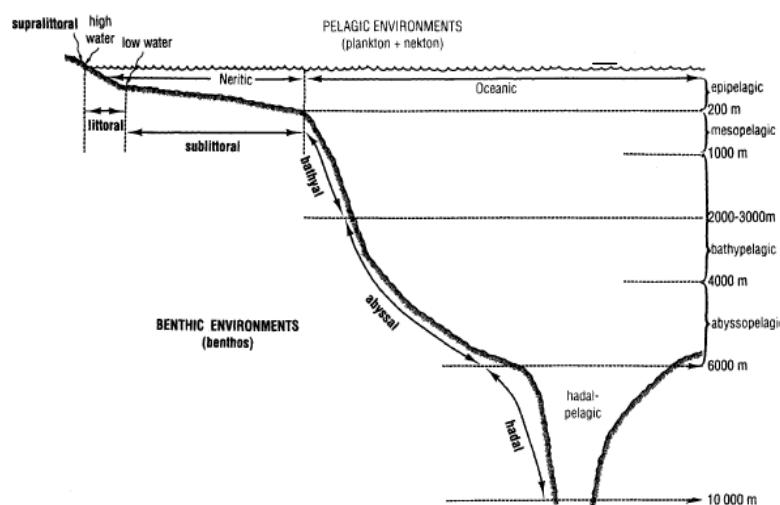


Fig. 4.23 vertical zones as per light penetration

The Fig.4.22 shows the relationship between Light Intensity (L.I) and productivity at different depths. Both the plant and animal communities depend on the availability of light intensity in the under water. But note that although the light intensity is maximum at the surface (0 m), the maximum productivity is not confined at the surface but concentrated around 20 to 30 m depth. In any case entire productive zone is in the upper 100 m or so. The quick absorption of light by the waters in different depths is seen. when it goes to 15 m, light intensity (L.I) becomes half (50%), at 25 m the availability of light is only 1/3 (36%), when it goes to 35 m or so it is only 1/5 (20%) and at 50 m, the light availability is 1/10 (10%). Which means 90% is absorbed by the time light penetrates to 50 m or so.

Fig.4.23 shows the occurrence of different zones in the sea in the vertical as per the penetration of sun light in the vertical in the oceans which is commonly used by marine biologists. The depth zone between 0-200 m is called epipelagic, photic or sunlit zone as most of the sun light reaches in the zone. 200-1000 m is called meso-pelagic zone where the light penetration is poor and so it is dim light (twilight) zone. 1000-4000 m is called bathypelagic zone. Sun light doesn't reach into this zone and so it is called midnight zone. 4000-6000 m is called abyssal pelagic zone. The word abyssal in greek terminology means no bottom. Greater than 6000m is called hadal zone. It is in the trenches.

4.7.4. VARIATION OF 'K' IN DIFFERENT WATERS:

The variation of extinction coefficient of solar radiation with wave length in pure water and in different types of water is shown in Fig.4.24. The selective extinction of day light in the whole range of the visible part of the spectrum shows remarkable differences with different water types. With pure water the extinction is the strongest in the range of longer wave length.

With increasing turbidity of the open sea water, and especially of coastal water, the extinction in the short wave length (blue) (420-480 m μ) becomes more significant. The selective extinction produces a shift of the minimum extinction coefficient from shorter to longer wave lengths as the turbidity increases. Clear open ocean water is most transparent for blue light of about 470 to 480 m μ whereas in turbid coastal water the maximum transparency has shifted to about 550 m μ (the yellow-green).

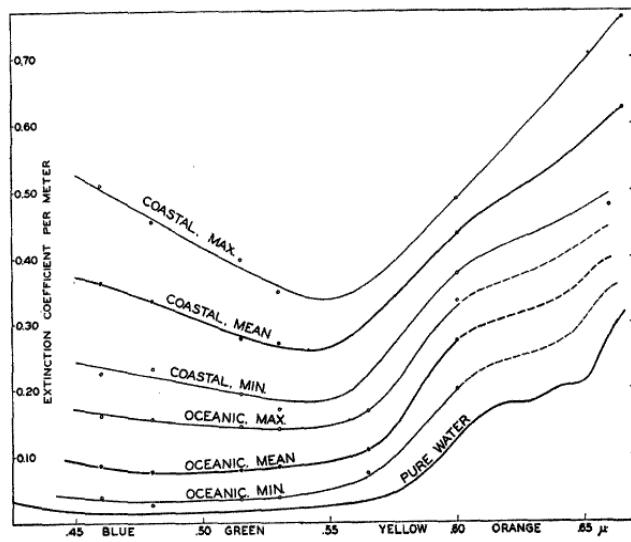


Fig.4.24 Variation of extinction coefficient in different types of waters (after Sverdrup et al 1942)

4.7.5. COLOR OF SEA WATER:

The blue color of sea water is due to Raley's scattering at the sea surface and the substances dissolved and suspended in different areas will change the color from yellowish green to deep blue color. While the largest part of the surface of the ocean in the tropics and sub tropics is blue, the coastal areas, shallow seas and Polar Regions are greenish in color.

In deep water if one looks downward below the surface, the light scattered is due to the molecules of water. Because the water molecules scatter the short wave (blue) of visible spectrum much more than the long wave (Red) band, the color seen is selectively blue. In addition the red and yellow wave lengths of sun light are rapidly absorbed by the upper few meters and the only light remained to be scattered in the deeper water is blue. If the sky is blue during the cloud free time the blue color is enhanced and so appears deep blue, but if the sky has clouds they will dilute the blue color and so it appears pale blue.

If there is green phytoplankton their chlorophyll content will absorb the blue light and shift the water color to green. In coastal regions and river mouth areas, rivers bring large sediment containing organic substances, so coastal waters appear brown or yellowish green color.

4.8. ACOUSTICAL PROPERTIES OF SEA WATER

Oceanographers map the sea floor features by using an echo- sounder. It emits sound pulses from the ship towards the bottom from where they are reflected to the surface, and time taken is recorded. Knowing the velocity of sound in ocean water, they find out the water depth below the ship. Present day multi-beam echo sounders provide wide coverage of the sea-floor depth simultaneously. Side scan sonar deployed at shallow water depths gives a picture of the features of the sea-floor and sunken ships and similar objects. Sound of different frequencies can penetrate through the sea-floor, get reflected and come back to the ship. Through, such seismic investigations, scientists are able to find out structure and materials lying beneath the sea-floor. This is how sound is useful in scientific study.

Acoustic tomography has been used to investigate fronts and to identify and track mesoscale eddies. The method relies on the fact that individual eddies have temperatures different from those in the surrounding water- there are both warm and cold eddies. It follows that the speed of sound between an acoustic source and receiver will change if an eddy passes between them.

Sound may be used to measure current velocity by exploiting the Doppler Effect whereby the measured sound frequency is affected by relative movement between an acoustic source and the point of measurement. Narrow-beam echo-sounders are aimed at a particular volume of water and the shift in frequency between the sound waves emitted by the hydrophones and those back-scattered by particles in the water is measured. This Doppler shift in frequency is proportional to the current speed, which can therefore be determined. Thanks to rapid technological advances, since the late 1980s it has become routine practice for ship-mounted Acoustic Doppler Current Profilers (ADCPs) to be used for continuous measurement of current velocities to depths of several hundred metres, while the ship is actually in motion.

SOFAR (Sound Fixing and Ranging) is used for longer-range location, and also for tracking, especially of neutrally buoyant acoustic floats within and near the sound channel. To fix the position of Sofar devices reliably, variations of the speed of sound throughout the oceans must be known as accurately as possible. The axis of the sound channel lies between about 0.5 and 1.5 km depth throughout most of the oceans, between the latitudes of about 60° N and S. Poleward of these latitudes there is no sound channel.

Sound is a vibration composed of frequency that travels through solids, liquids, & gases and capable of being detected by ears. We cannot hear all the sounds. The audible range of human ear is between 20 Hz to 20,000 Hz. The sound below 20 Hz is called infrasonic whereas beyond 20,000 Hz it is called ultrasonic. The branch of science that deals with the study of sounds is called, acoustics.

Loudness, pitch and quality are the three characteristics of sound. The loudness of sound depends upon the amplitude of vibration. The pitch is determined by frequency of vibration. The quality of sound depends on vibrating body. The loudness of sound is measured in decibels.

4.8.1. INTENSITY OF SOUND:

The intensity 'I' of the sound is proportional to the square of the pressure fluctuation 'P':

$$I = \frac{P^2}{\rho C}$$

Where 'ρ' is the average density of sea water and 'C' is the velocity of sound. Intensity has units of energy/area-time. It measures the flux of sound energy past a point and the energy travels with the speed of sound.

Sound pressure varies between less than 10^{-4} dynes/cm² to more than 10^6 dynes/cm². The lower limit is less than the audible sound and the higher limit is equal to the atmospheric pressure.

4.8.2. SOUND LEVEL (L) AND DECIBEL (dB):

Originally used for measuring sound amplitudes, the bel and decibel are also used for measuring amplification and abatement. The Bel unit is named after the inventor of the telephone, Alexander Graham Bell (1847-1922). One Bell louder means ten times louder, measured by amplitude.

1 bel = 1 B = $\log (I / I_0) = 10$ decibels ; where I = current intensity and I_0 is previous intensity, or a sound level of 2E-10 bar (2×10^{-10}). Please note on the Sea friends web site, the powers of ten are often written as 1E6 rather than 10^6 or 10^6 . For example, $15000 = 15E3 = 1.5E4$ or $15E-3 = 15 \times 10^{-3}$.

$$1 \text{ decibel} = 1 \text{ dB} = 10 \times \log(I / I_0)$$

The range of human hearing extends from 0 - 120 dB, at which level sound becomes painful, while also causing permanent deafness. The critical level for the start of ear damage is 85 dB. Conversation: 30-60 dB; jet plane 160 dB; Noise from trucks can exceed 80 dB measured at the roadside. People begin to complain when sound levels in their neighbourhood reach 35-40 dB.

Thus sound intensity is usually discussed in terms of a non dimensional quantity called decibel (dB). Sound level in decibels is ten times of the logarithm to the base 10 of the ratio of the measured sound intensity to the reference intensity as mentioned above.

$$\frac{I}{I_r} = \frac{P^2 / \rho C}{P_r^2 / \rho C} = \frac{P^2}{P_r^2}$$
$$10 \log\left(\frac{I}{I_r}\right) = 10 \log\left(\frac{P^2}{P_r^2}\right)$$

For example, for a sound intensity with a sound pressure of 1200 dynes/cm² with reference pressure of 1 dyne/cm², the sound level is

$$\log\left(\frac{I}{I_r}\right) = 10 \times 6.16 \text{ bels} = 61.6 \text{ dB} = 62 \text{ dB}$$

Thus for a sound intensity with a sound pressure of 1200 dynes/cm², the sound level is 62 decibels.

4.8.3. SOUND FREQUENCY:

The number of vibrations extended by a vibrating particle of a medium about its mean position in one second is called its frequency. Its unit is hertz (Hz) in S.I units. One hertz is one wave per second. Also one hertz is 1 cycle/second = $2 \pi \text{ rad/s} = 6.28318 \text{ rad/s}$

Sound pitch is usually measured in terms of frequency. Human ear can detect as high as 15000 to 18000 Hz and as low as 20 to 50 Hz. If the velocity of sound is known, the wave length at different frequencies can be detected from the equation $V = n\lambda$.

Assuming a sound velocity of 1500 m/s typical wave lengths for under water acoustics are as follows:

Frequency	10^2	10^3	10^4	10^5
Wave length	15 m	1.5 m	0.15 m	0.015 m

Most echo sounders and SONAR operate at frequencies in this range. Echo sounder is a device used to determine the depth of a shoal or a hidden submarine. It contains a sound source and a pair of hydrophones. After sending the sound pulse vertically downward into the sea, when it touches a hidden shoal, submarine or a school of fish it gives an echo. The echo can be heard either by hydrophones or can be recorded on an echogram.

A 'SONAR' is Sound Navigation And Ranging which is an advanced equipment and very valuable for submarine war fare and other navigational purposes.

4.8. 4. HISTORY ON UNDER WATER ACOUSTICS:

Underwater sound has probably been used by marine animals for millions of years. Aristotle is thought to have been the first who noted that sound could be heard in water. Almost 2000 years later, Leonardo da Vinci wrote, "If you cause your ship to stop and place the head of a long tube in the water and place the other extremity to your ear, you will hear ships at great distances." In 1687 Isaac Newton wrote his Mathematical Principles of Natural Philosophy which included the first mathematical treatment of sound.

The modern study of underwater acoustics can be considered to have started in early 19th century. In 1826, in Lake Geneva, the speed of sound was first measured (Fig 4.25) by Daniel Colladon, a Swiss physicist, and Charles Sturm, a French mathematician. Their calculation of the speed of sound (1435 m/s) was only three meters less than presently accepted values.

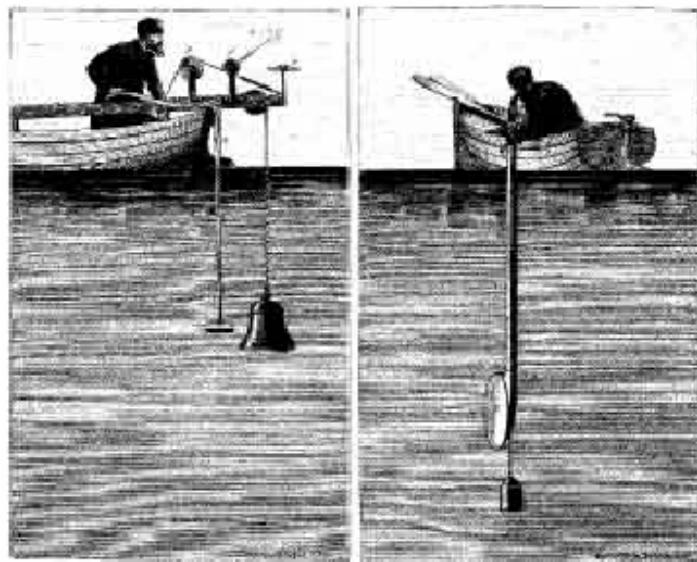


Fig.4.25 Measuring the speed of sound in Lake Geneva in Sept 1826

Later, in the 1890s, Elisha Gray, designed a waterproof telephone transmitter that could be used as a “hydrophone” to listen the underwater bell signals. Shortly after, in 1912, the effort was expanded to include underwater communication through transmission of Morse code in the sea. The sinking of the Titanic in 1912 and the start of World War I provided the impetus for the next wave of progress in underwater acoustics.

The advent of World War I and advances in submarine warfare drove the development of underwater acoustic technologies. Echoes were received from a submarine at distances as great as 1500 meters for the first time, in 1918. This resulted in several new inventions, including the fathometer, and seismic prospecting. In years between the two World Wars, the development of underwater acoustic instrumentation benefited from advances in electronics, which allowed for amplification, processing, and the display of acoustical data. It was during time that scientists began to understand how sound propagates in the sea.

By the time World War II began, many American ships were equipped for both underwater listening and echo-ranging. During World War II, an increased effort in undersea acoustics was directed toward developing systems to locate and track German U-boats. Scanning sonar sets, acoustic mines, the acoustic homing torpedo, and nonreflecting coatings for submarines were all developed during wartime.

At the end of World War II, the Soviets gathered the German resources they needed to build up a submarine fleet, creating a new threat for the allied forces. Therefore, the US and British efforts in underwater acoustics continued and as a result, great advances in sound propagation were made. Sonar systems grew larger, more powerful, and operated at lower frequencies, resulting in much greater range.

Since the end of the Cold War, advances in underwater acoustics have continued. The focus of military acoustics has shifted from deep water to the shallow water. Moreover, a considerable increase in the use of underwater sound for non-military, commercial applications

has occurred. These commercial devices include side-scan SONARs to image ship wrecks, sub-bottom profilers to penetrate the seafloor while searching for oil and other minerals, acoustic speedometers for measuring ship speed, acoustic transponders and beacons for position marking, and myriad devices to aid in ocean exploration.

The greater part of ocean is almost completely dark and any artificial light that is introduced is subject to scattering and absorption making visibility poor. Sound propagates well through liquids and solids, less well in gases and not at all in vacuum. Acoustic waves, however, travel well in the ocean and this makes possible the remote sensing of objects and the transmission of information.

The speed of sound is affected by the oceanographic variables of temperature, salinity, and pressure. Sound speed increases with increasing pressure (depth), temperature and salinity. A typical speed of sound in water near the ocean surface is about 1520 meters per second which is more than 4 times faster than the speed of sound in air. The approximate change in the speed of sound with a change in each property is given below:

$$\text{Change in Temperature by } 1^{\circ}\text{C} = 4.0 \text{ m/s}$$

$$\text{Change in Salinity by 1 PSU} = 1.4 \text{ m/s}$$

$$\text{Change in Depth (pressure) by 1000 m} = 17 \text{ m/s.}$$

In the top few hundred meters of ocean, where temperature changes are large, sound speed will be controlled mainly by temperature and to a much smaller degree by salinity and depth. Below the permanent thermocline, neither T nor S varies greatly and so pressure becomes dominant control on sound speed. Horizontal variations in sound speed are very much smaller than vertical ones because horizontal gradients of temperature and salinity are much smaller than vertical gradients. Thus, an acoustic wave traveling vertically in the ocean will not be significantly affected by refraction because it is traveling essentially at right angles to the interfaces between layers of different density. However, the wave traveling horizontally may undergo considerable refraction because it will meet such interfaces at low angles.

The vertical profile of sound speed in the ocean is given in Fig.4.26. The decrease in sound speed near the surface is due to decreasing temperature. The sound speed at the surface is fast because the temperature is high from the sun warming the upper layers of the ocean. As the depth increases, the temperature gets colder and colder until it reaches a nearly constant value. Since the temperature is now constant, the pressure of the water has the largest effect on sound speed. Because pressure increases with depth, sound speed increases with depth. Salinity has a much smaller effect on sound speed than temperature or pressure. In all water sound speed is determined by its bulk modulus and mass density. The bulk modulus is affected by temperature, dissolved impurities (usually salinity), and pressure.

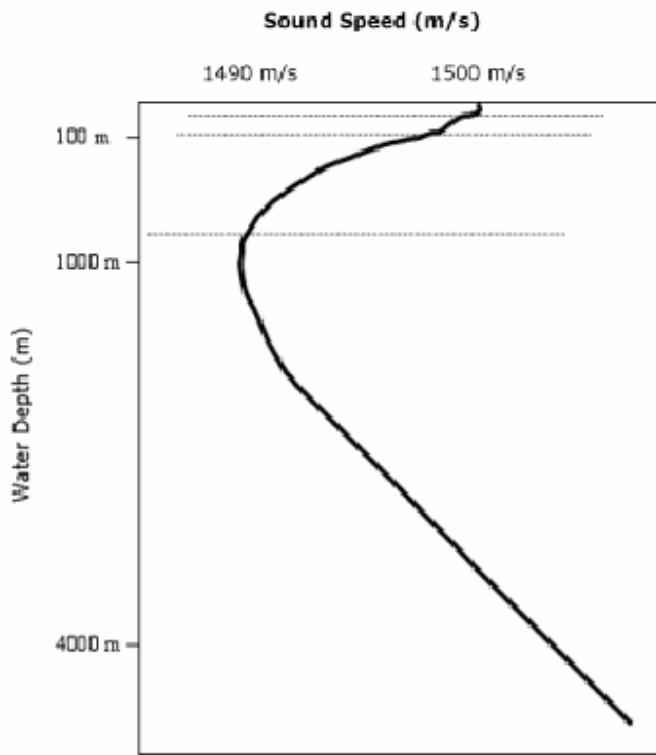


Fig.4.26. Typical sound speed (m/s) profile in the oceans

The sound Frequency range used in the oceans is 1 Hz (1 cycle/second) to 100 KHz. The important applications of sound in the sea are submarine signalling, communications, detection, acoustic Doppler Current Profiler, bathymetric measurements, planktonic & fish detection (deep scattering layer), upper ocean wave breaking & turbulence (bubbles) and large scale temperature monitoring for Climate.

4.8.5. THE PROPAGATION OF SOUND IN SEA WATER:

Although both light and sound travel through water, they travel in different ways. While light travels as a transverse wave, sound travels as a longitudinal wave. This means that in the case of light the oscillation amplitude is at right angles to the direction of propagation whereas sound propagates as a pressure disturbance with compression and rarefaction occurring alternately in the direction of propagation.

4.8.6. ACOUSTIC EQUATION:

It is derived from momentum equation and from hydrodynamic continuity equation. Linear equation is used for small amplitude pressure and velocity signals. This equation predicts “ray paths” of sound waves.

$$\frac{\partial^2 p}{\partial t^2} = \frac{\kappa}{\rho_o} \left(\frac{\partial^2 p}{\partial x^2} + \frac{\partial^2 p}{\partial y^2} + \frac{\partial^2 p}{\partial z^2} \right)$$

$$C = \sqrt{\frac{\kappa}{\rho_o}} \text{ where } \kappa = \text{bulk modulus} = \rho_o \frac{\partial \rho}{\partial p}$$

Where C is sound velocity, p is pressure and ρ density.

In the ocean, the sound process is “adiabatic” and so very rapid compressions and rare factions occur. Therefore very little heat is lost to surroundings due to propagation of sound in the ocean.

The transmission of sound with low attenuation in the ocean enables scientists to use acoustical methods to determine the depth of the oceans, sea bottom structure and to locate the hidden submarines as well as to communicate over considerable distances in a medium that is nearly opaque to most electromagnetic radiation. That is how all the marine animals can use sound as a communicating device.

Water transmits sound more effectively and faster than the atmosphere. Another way of the equation of velocity of sound is given by

$$C = \sqrt{\frac{\gamma}{\rho k}}. \dots \dots \dots (10)$$

Where γ is ratio of specific heats (C_p/C_v), ρ is density of water and ‘k’ is the compressibility (or $1/k$ is called as the elasticity of the medium). It is evident that an easily compressible material has a small elasticity and so velocity of propagation in such a material is low.

From the equation (10) it is clear that as elasticity is more, velocity is more. Elasticity will be more if the material is incompressible, because the former is the reciprocal of the latter ($1/k$). So for all practical purposes water is considered incompressible.

As elasticity depends on temperature, salinity and σ_t , sound velocity (V) also depends on all the three parameters in the oceans. In the oceans the velocity of sound increases with increasing salinity, temperature and pressure. In the upper layers sound velocity depends on variation of temperature and in deeper layers it depends on pressure as the variation of temperature and salinity are less there.

Sound needs a medium to travel. It propagates best through solids and liquids and less well in gases but not at all in vacuum. The speed of sound is not the same in solids, liquids and gases. The speeds of sound in some media are given in Table 4.3.

Medium	Speed of sound (m/s)
Air	343
Fresh Water	1440
Sea water	1500
Brick	3542
Steel	5100
Glass	5000 - 6000

Table 4.3. Speed of sound in different media

Sound travels 333 m/s in air but in sea water it travels at a speed of 1445 m/s which is nearly 4 to 5 times faster than in the atmosphere. The sound velocity does not change much with normal variations of density and salinity in the oceans but it changes much with variation of temperature.

Fig.4.27 shows the effect of sea pressure (depth) on sound velocity and also the comparison of variation of sound velocity and temperature. This effect is almost independent of temperature and salinity. Thus the sound velocity increases i) at a rate of 4-8 m/s for every 1°C increase of temperature ii) at a rate of 1 m/s for every 1‰ increase of salinity and iii) at a rate of 2 m/s for every 100 m increase of depth. The variations are more rapid in the vertical than in the horizontal direction in the oceans.

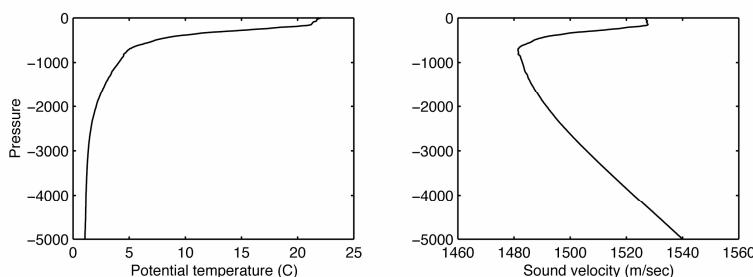


Fig. 4.27 Relation between sea pressure, temperature and sound velocity

4.8.7. Absorption by sea water:

While the sound is propagating some amount is absorbed by waters which is given as

$$\frac{dI}{dR} = -jI$$

$$I = I_o \exp(-jR)$$

Where I is acoustic intensity Wm^{-2} , j is the acoustic absorption coefficient m^{-1} and R is the distance.

Typical values of j (in decibels, dB, per kilometer) are: 0.08 dB/km at 1000 Hz, and 50 dB/km at 100,000 Hz. Decibels are calculated from: $\text{dB} = 10 \log (I/I_0)$, where I_0 is the original acoustic power, I is the acoustic power after absorption.

For example, at a range of 1 km a 1000 Hz signal is attenuated by only 1.8%: $I = 0.982I_0$. At a range of 1 km a 100,000 Hz signal is reduced to $I = 10^{-5}I_0$. The 30,000 Hz signal used by typical echo sounders to map the ocean's depths is little attenuated going from the surface to the bottom and back.

4.8.8. SOUND VELOCITY EQUATION USED IN THE OCEANS:

The relationship for sound velocity in sea water as a function of temperature and salinity is given by the equation:

$$C = 1449 + 4.6 t - 0.055 t^2 + 0.0003 t^3 + (1.39 - 0.012 t) (S - 35) + 0.017 d.$$

Where C is sound velocity, t is temperature in degrees Celsius, S is salinity and 'd' is depth in decibars (meters) of the oceans.

Ready made tables are available for calculation of sound velocity if t , s and σ_t are known. These tables were prepared using the above formula.

The sound speed c is a function of temperature T , salinity S and pressure p and varies between 1400 m s^{-1} and 1600 m s^{-1} . In the open ocean it is influenced by the distribution of T and p but not much by S . It decreases with decreasing T , p and S . The combination of the variation of these three parameters with depth produces a vertical sound speed profile with a marked sound speed minimum at intermediate depth: Temperature decreases rapidly in the upper kilometre of the ocean and dominates the sound speed profile, i.e. c decreases with depth. In the deeper regions (below the top kilometre or so) the temperature change with depth is small and c is determined by the pressure increase with depth, i.e. c increases with depth. Vertical changes of salinity are too small to have an impact; but the average salinity determines whether c is low (if the average salinity is low) or high (if the average salinity is high) on average.

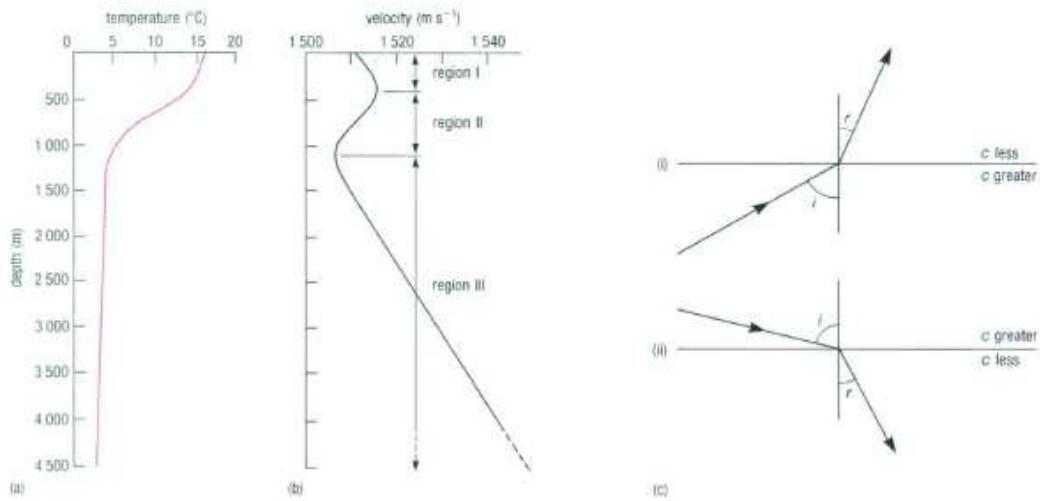


Fig.4. 28 (a) A typical temperature profile in the ocean. (b) A typical profile of the speed of sound in the ocean, (c) Idealized sketches illustrating refraction at interfaces where the speed of sound changes differently

Figure 4.28(b) shows the variation of sound speed in different depth zones of the oceans. The speed of sound is shown to increase with depth in the mixed layer, region I because T and S are more or less constant there and so c is controlled chiefly by pressure. Region II coincides with the permanent thermocline (and main halocline), where c is controlled chiefly by changes in T and S. Below the permanent thermocline, in region III, the speed of sound is almost entirely controlled by pressure.

Fig.4.28(c) shows the refraction at interfaces where the speed of sound changes. (i) Upward refraction (regions I and III) and (ii) downward refraction (region II). From Snell's law:

$$\frac{C_{\text{greater}}}{C_{\text{less}}} = \frac{\text{Sini}}{\text{Sinr}}$$

In regions I and III of Figure 4.28(b), a sound wave will be refracted upwards, because the speed of sound decreases upwards as in Figure 4.28(c) whereas in region II it will be refracted downwards, because the speed of sound decreases downwards as in Figure 4.28 (c).

Figure 4.28 (d) is an example of a ray diagram for a sound emitted in region II of Figure 4.28(b), showing a sound channel and a shadow zone. The shadow zone is defined by the limiting rays, reflected at the sea-surface and refracted at the boundary between regions I and II. The rays are simply lines drawn perpendicular to the propagating wave front, and they therefore represent the direction of propagation. Note that most rays are focused on the boundary between regions II and III, whereas there is a shadow zone in the vicinity of the boundary between regions I and II that is penetrated only by waves that have been reflected at the surface of the ocean. The channel in which rays are trapped by refraction at the boundary between regions II and III is known as the sound channel, which is in effect a 'wave guide' for sound in the oceans.

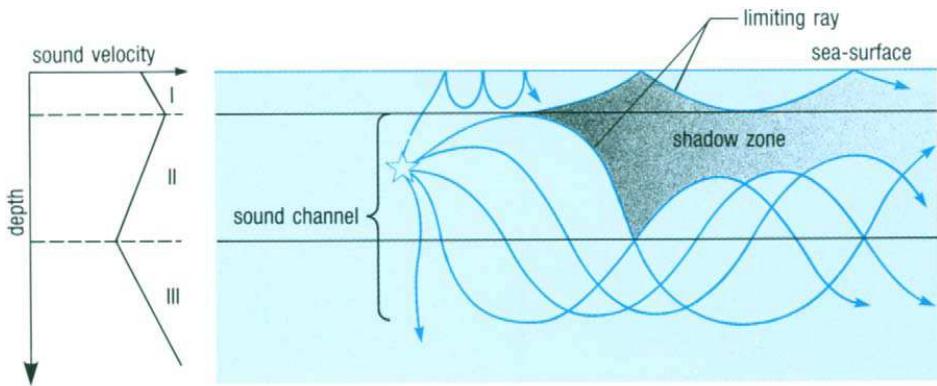


Figure 4.28 d: An example of a ray diagram for a sound emitted in region II of Figure 4.28(b), showing a sound channel and a shadow zone.

4.8.9. Sound velocity in the Arabian Sea and the Bay of Bengal:

As discussed earlier, the characteristics of the sound speed profile arise from the characteristics of the temperature and salinity profiles (Fig.4.29) of the locality. In both the seas of Arabian Sea and bay of Bengal the temperature minimum is at 300 m and so the minimum sound velocity also occurs at 300m (Fig.4.30).

The vertical profile of sound speed in the Arabian Sea shows that in the upper 50 m the sound speed increases with depth reaching a maximum value of about 1548.8 m/s at 50 m and below this the sound speed shows a steady decrease reaching the minimum value of 1540.2 m/s at 300m depth (Fig.4.30 left). Below this depth the sound speed shows an increase right up to the bottom. Thus the characteristics of the sound speed profile in the Arabian Sea is the presence of surface duct in the upper 50 m which is very important for the submarine SONAR operations. The SOFAR channel axis is at 300 m and the effective wave guide lies between 100 m and 1250 m. The surface duct seen in the sound speed profile in the upper 50 m in the Arabian Sea is due to the presence of isothermal and isohaline layers within which the increase in depth leads to an increase in the sound speed, giving rise to the observed surface duct. The decrease in sound speed below the surface duct and up to 300 m is due to the rapid decrease of temperature and salinity (thermo cline and halocline) in this layer. Below 300 m the effect of increased pressure due to increase in depth is more dominant than the slow decrease of temperature and salinity with depth which ultimately increases the sound speed.

The vertical sound speed profile in the Bay of Bengal shows an increase of sound speed from 1544 m/s at the surface to about 1546 m/s at about 60 m (Fig.4.30 right). Below this the sound speed shows rapid decrease and reaches the minimum value of 1537 m/s at 300 m. Below 300 m the sound speed increases steadily right up to the bottom. Thus, the characteristic feature of the sound speed profile in the Bay of Bengal is the presence of surface duct in the upper 60 m and the SOFAR channel axis is at 300m. The effective deep wave guide is between the depths 100 and 1200 m. The strong surface duct seen in the vertical profile of sound speed in the Bay of Bengal arises due to the sharp increase of salinity in the isothermal layer. Thus, in the Bay of Bengal both increasing salinity and depth gives rise to the formation of strong surface duct. As in the case of the Arabian Sea, the increase in sound speed below 300 m is due to the increase in pressure which has more effect in the increase of sound speed compared to the slowly decreasing temperature and salinity.

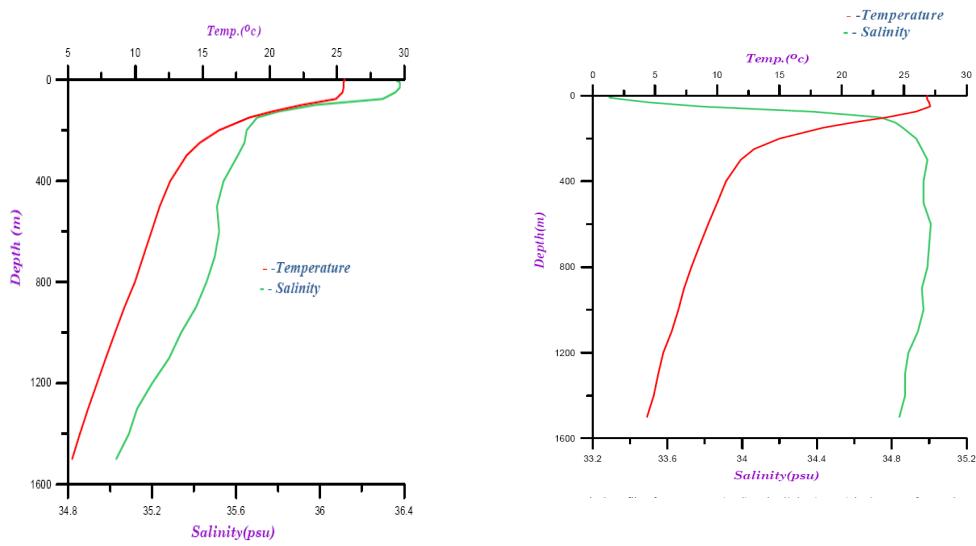


Fig.4.29. Vertical profile of temperature(red) and salinity(green) in the Arabian sea at 14.5°N,63.5°E (**Left**) and Bay of Bengal 14.5°N,87.5°E (**Right**)

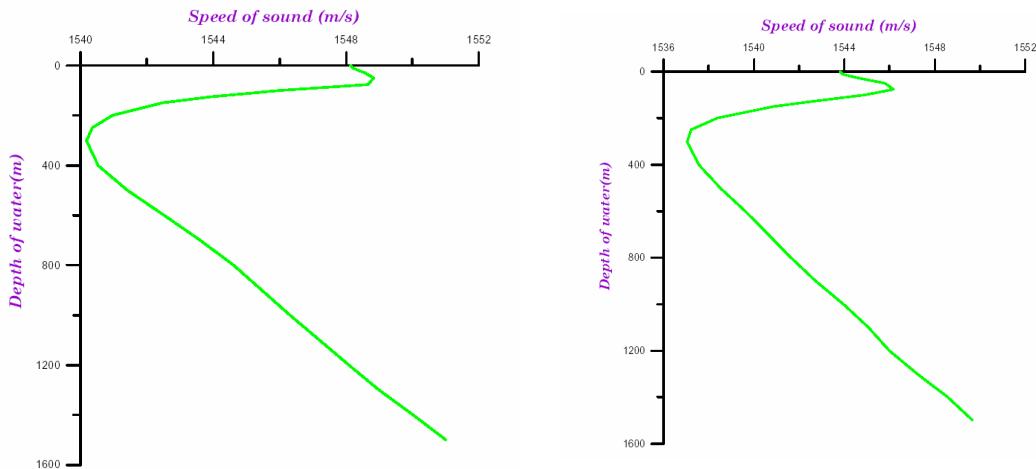


Fig.4. 30 Vertical profile of sound speed in Arabian sea at 14.5°N, 63.5°E (**Left**) and Bay of Bengal at 14.5°N,87.5°E (**right**)

4.8.10. DEPTH DETERMINATION:

The sound velocity can be used to determine the depth of a shoal, lurking enemy submarine or a school of fish. As there are different zones in the oceans, sound travels at different speeds. Normally the sound velocity profile follows the temperature profile in the vertical. So in the mixed layer it is constant, thermocline it decreases to maximum and later the decrease is slight. Because of these differing velocities in different zones, a weighted average of sound velocity is used to determine the depth from the equation:

$$D = (\frac{1}{2}) t.v$$

where 't' is the time taken for the sound pulse to travel from the ship to the sea bottom and back to the ship.

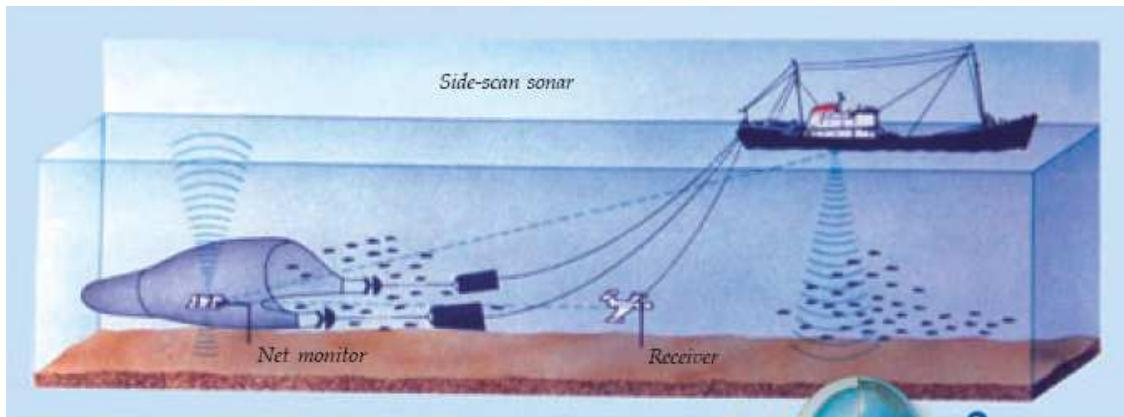


Fig.4. 31a .Mapping the sea floor or identifying school of fish using side scan sonar

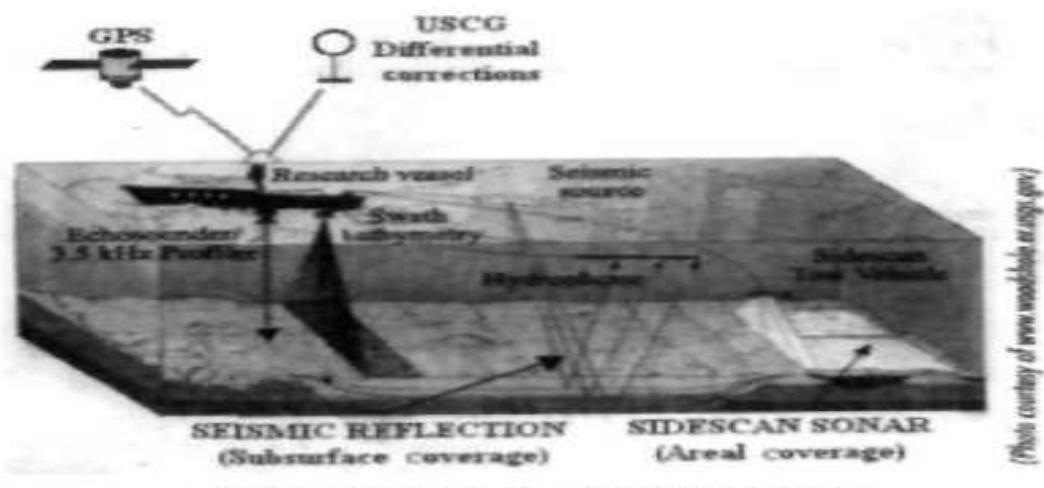


Fig.4. 31b. Mapping the sea floor using multibeam echo-sounder (swath bathymetry)

Oceanographers map the sea-floor features by using an *echo sounder or side scan sonar* (Fig.4.31a). It emits sound pulses from the ship towards the bottom from where they are reflected to the surface, and the time taken is recorded. Knowing the velocity of sound in ocean water, they find out the water depth below the ship. Present day multi-beam echo sounders provide wide

coverage of the sea-floor depths simultaneously. The side-scan *sonar* deployed at shallow water depths gives a picture of the features of the sea-floor and sunken ships and similar objects.

What lies beneath the sea-floor? Sound of different frequencies can penetrate through the sea-floor, get reflected and come back to the ship. Through such *seismic investigations*, scientists are able to figure out what materials are there below and what their structures are.

Multibeam echo sounder (Fig.4.31b) is a relatively recent successor to echo sounding technique. This technology utilizes multiple beams of sound at the same time to cover a large fan-shaped area of the ocean floor. This is in contrast to the technique of echo sounder in which a small patch of seafloor is covered and mapped. Multibeam echo sounder sends out a swath of acoustic beams that covers a distance on either side of the ship. All of the signals that are sent out reach the seafloor and return at slightly different times. These signals are received and continuously converted through software to water depths, and then automatically plotted as bathymetric maps.

4.8.11. WORKING PRINCIPLE OF ECHOSOUNDER:

The instrument transmits a burst of 10–30 kHz sound and listens for the echo from the sea floor. The time interval between transmission of the pulse and reception of the echo, when multiplied by the velocity of sound, gives twice the depth of the ocean (figure 4.31c).

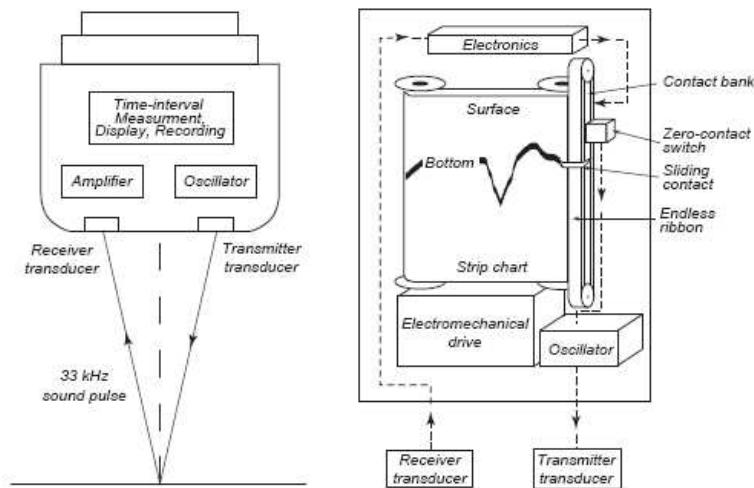


Fig.4. 31c. **Left:** Echo sounders measure depth of the ocean by transmitting pulses of sound and observing the time required to receive the echo from the bottom. **Right:** The time is recorded by a spark burning a mark on a slowly moving roll of paper. After Dietrich et al (1980: 124)

4.8.12. SOUND REFRACTION:

Sound waves beamed obliquely over the surface reflect because of differing zones of varying temperature, salinity and density and so it obeys Snell's law. As the sound beam travels from zone I to zone II, it suffers refraction as shown in Fig.4.32. Let V_1 is the incident velocity and V_2 is the refracted velocity. By the time the refracted ray at 'O' reaches Q ($V_2 \Delta t$), the non refracted

second ray of the beam in zone I reaches from A to B ($V_1\Delta t$), Where Δt is the time difference between the first and second ray after and before refraction respectively.

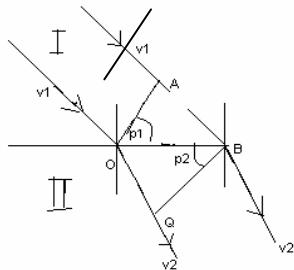


Fig.4.32 .Sound refraction

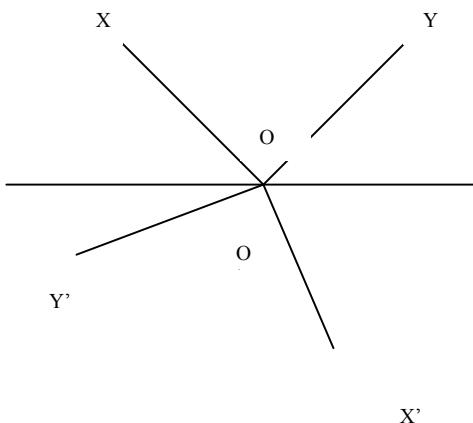


Fig.4. 33 Sound cones formed by refraction

$$\text{Therefore from right angled triangles OAB and OQB } \frac{\sin p_1}{\sin p_2} = \frac{V_1 \Delta t}{V_2 \Delta t} = \frac{V_1}{V_2} \text{ where } p_1 \text{ and } p_2 \text{ are angles of incidence and refraction.}$$

and p_2 are angles of incidence and refraction. As here v_2 is greater than v_1 , downward bending results. The wave fronts are perpendicular to the rays which show the direction of propagation. In some cases, if angle p_1 is large enough, the ray does not penetrate the boundary and total reflection occurs thus p_1 is equal to or greater than the critical angle.

In fig.4.33 XOY and X'YO' are cones of sound before and after refraction respectively. If a cone of sound passes through water of uniform sound velocity (mixed layer) and then strikes a zone in which the sound speed is smaller (thermocline) as shown in Fig.4.34b, then the refracted cone is bent downward. Similarly if the sound passes from lower velocity zone to higher velocity zone, it bends upward as shown in Fig.4.34a.

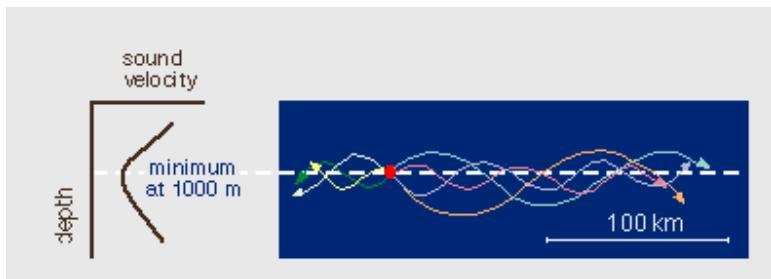


Fig.4. 34a Horizontal sound paths in the SOFAR channel (top) and near the surface, if the surface layer is well mixed (bottom). The diagram at the left gives the vertical sound velocity profile; note the profile shape. The red circle indicates the sound source. (By courtesy of www.incois.gov.in/Tutor/IntroOc/notes/figures/fig5a4..html).

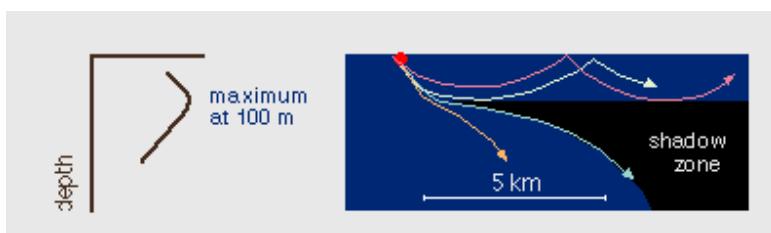


Fig.4. 34b Horizontal sound paths in the SOFAR channel, if the surface layer is well mixed. The diagram at the left gives the vertical sound velocity profile; note the profile shape. The red circle indicates the sound source. (By courtesy of www.incois.gov.in/Tutor/IntroOc/notes/figures/fig5a4..html).

Fig.4.34. gives examples of horizontal sound paths. The first diagram (Fig4.34a) shows sound propagation at the depth of the sound speed minimum (usually about 1000 m). Sound rays bend back towards the depth of minimum sound speed and travel at that depth over large distances (they can traverse entire oceans). This sound channel is known as the SOFAR (SOund Fixing And Ranging) channel. Before the introduction of the Global Positioning System (GPS) the SOFAR channel was used to locate ships and aircraft in distress, and for tracking floats (with two or more receivers) for the study of ocean currents. The second diagram (Fig.4.34b) shows a situation where a mixed layer of uniform temperature (typically about 100 m thick) is found on top of the normal temperature stratification. In this case sound speed increases below the surface due to the increase in pressure before the normal decrease due to temperature takes over. The resulting sound speed maximum at about 100 m depth creates a shadow zone, since all sound rays bend away from that depth. The principle of sound propagation is that sound rays always bend towards the region of lower sound speed. This produces a sound channel near 1000 m depth (top) and a shadow zone below the surface mixed layer (bottom) to which sound from the source cannot penetrate. Since sound rays are reversible, this also means that sound produced in the shadow zone cannot be heard by a sound detector placed at the location as shown by red spot.

Sound travels along a straight path where the sound speed c is constant; otherwise it bends toward the region of lower c . Different rays are independent of each other. Sound paths are reversible. The law of reflection (angle of incidence = angle of reflection) holds at the sea floor, surface, at objects and interfaces. The law of refraction holds at interfaces (Fig.4.35):

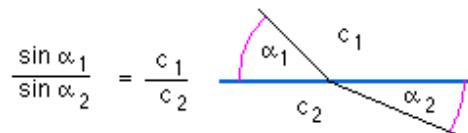


Fig.4. 35. Snells law holds good

As the stratification in the ocean is nearly horizontal, sound propagation in the vertical is practically along a straight path. This is the basis for echo sounding: The depth is known if the mean sound velocity is known. A first estimate is 1500 m s^{-1} ; available tables list corrections for the various areas of the world ocean.

Ray Tracing i)

Ray tracing (Fig.4.36) is the method by which the path of the sound can be determined in the ocean. It can show how sound channels are formed and why shadow zones occur. Based on Snell's law (Fig.4.35 or 4.37a) which relates sound velocity to direction of propagation α or ϕ with respect to horizontal

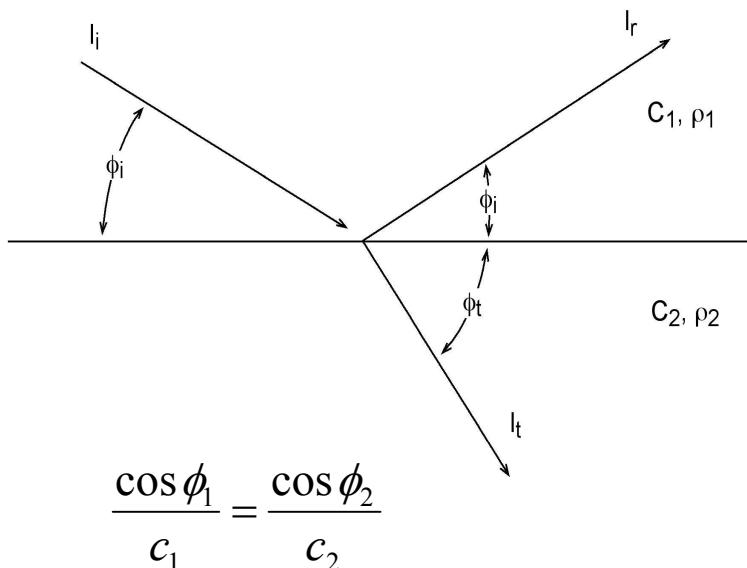


Fig.4. 36 Ray tracing method

The critical angle (ϕ_c) where no sound energy is refracted into higher velocity layer is shown in figure 4.37b.

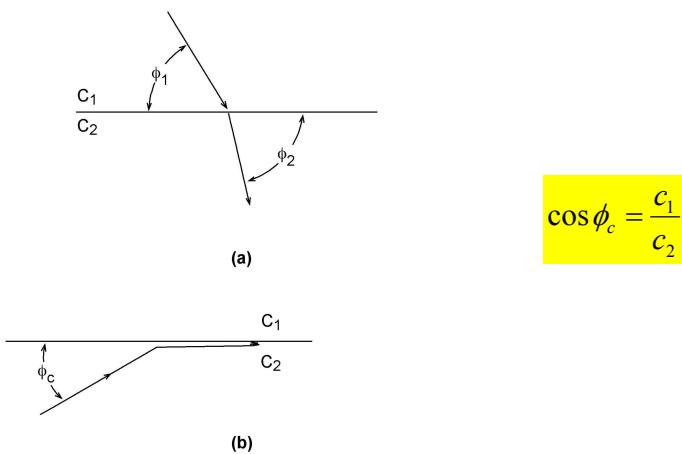


Fig.4. 37 the process of critical angle

For a constant vertical gradient in sound speed, dc/dz , the sound ray describes an arc with a radius, r_c as shown in Fig.4.38

$$r_c = \frac{c_o}{\left(\frac{dc}{dz}\right) \cos \phi_o}$$

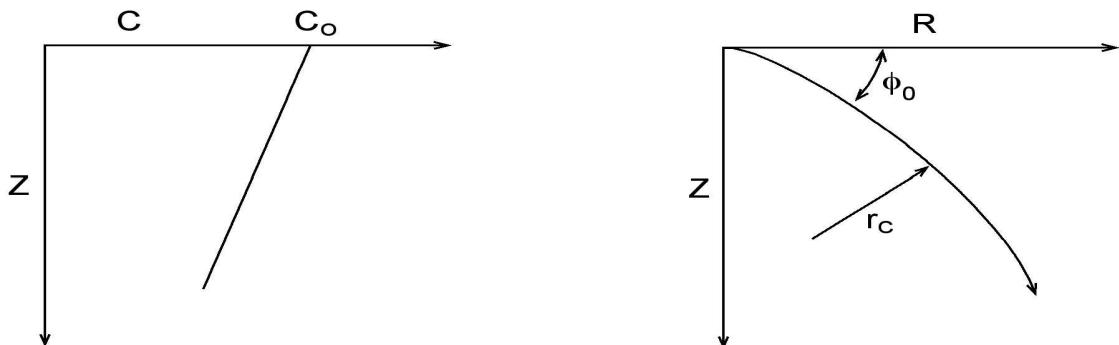


Fig.4. 38. The determination of radius of the arc of bending of sound ray

Reflection of Sound at Boundaries:

The reflection properties at boundaries depend on “acoustic impedance” (density \times sound velocity). For air-sea interface there is good reflection and poor transmission because acoustic impedance in air is low compared with water. (Water = 10^6 and air = 400). For water-sediment interface the ratio of the acoustic impedances is 1:3 hence high transmission occurs. From figure 4.34 we can write:

$$\frac{I_R}{I_i} = \left(\frac{\rho_2 c_2 \sin \phi_i - \rho_1 c_1 \sin \phi_t}{\rho_2 c_2 \sin \phi_i + \rho_1 c_1 \sin \phi_t} \right)^2$$

50

where I_R = reflected energy,

I_i = incident energy and I_t = transmitted energy

Acoustic Noise:

In the ocean when a sound pulse is sent with a certain frequency some ambient noise interferes with the source signal which depends on the background energy level. The sources of ambient noise are due to typical magnitude and frequency, scattering (redirecting sound in the wrong direction) and reverberation (self-generated noise).

4.8.13. Applications of Echo sounding:

The fact that sea water can transmit sound is of particular interest to the oceanographer. The transmission of sound with low attenuation enables scientists to use acoustical methods to determine depth, sea bottom structure and locate submerged submarines as well as to communicate over considerable distances in a medium that is nearly opaque to most electromagnetic radiation.

4.8.14. SONAR OPERATIONS:

The above principles of sound velocity, in the zones of thermocline and inversion layers of the ocean, are utilized for SONAR operations for submarine warfare.

There are regions in the sea that sound beams can not penetrate when they are directed at small angles called shadow zones. If an enemy submarine can identify and lurk in a shadow zone, it will escape from the attack. So the sound propagation in the oceans has considerable importance in military operations. Though the sound velocity follows the temperature profile, it doesn't go exactly like a straight path but follows a curved path. Because of this problem considerable error results in determining the target distance, so additional measurements of sound velocity propagation are to be made for this correction.

Sound is transmitted as a longitudinal wave in the form of pressure disturbance, with compression and rarefaction occurring alternately in the direction of propagation.

Warm water is less compressible than cold water, so sound speed is higher in warm water. Water at high pressure is less compressible than water at low pressure, so sound speed is higher at high pressure. These competing effects create a maximum sound speed at the sea surface (warm) and at great pressure, with a minimum sound speed in between. The sound speed minimum is an acoustic waveguide, called the "SOFAR" channel.

In the deep ocean, the slowest sound speed occurs at a depth of about 800 to 1000 meters, i.e. the region of the water column where the sound speed first decreases to a minimum value with depth and then increases, due to pressure. If an acoustic energy is emitted at the depth of the sound speed minimum, as the sound energy propagates upward the sound rays will be bent downwards due to refraction. Similarly as the sound energy moves downward from the region of minimum sound speed, the sound rays will be bent upward again due to refraction. This upward

and downward bending of rays below and above the sound speed minimum results in trapping the rays within the ocean and allows the propagation of sound energy horizontally over long distances within the ocean with out any energy loss. This makes the ocean a good sound channel or acoustic wave guide. This channel is called the SOund Fixing And Ranging, or SOFAR, channel.

SONAR

Sonar is the name given to the acoustic equivalent of radar. Sonar (Sound Navigation And Ranging) is the generic name of the technology that is used to locate objects underwater. Sonar systems are of two basic types - active and passive as in figure 4.39 . In active sonar, the system emits a pulse of sound and then the operator listens for echoes. In passive sonar, the operator listens to the sounds emitted by an object that one is trying to locate. In Sonar, sound signal is emitted and reflections are received from objects within the water (perhaps fish or submarines) or from the sea-bed. When the sound wave travels vertically down to the sea-bed and back, the time taken will provide a measure of depth of water if sound speed is also known.

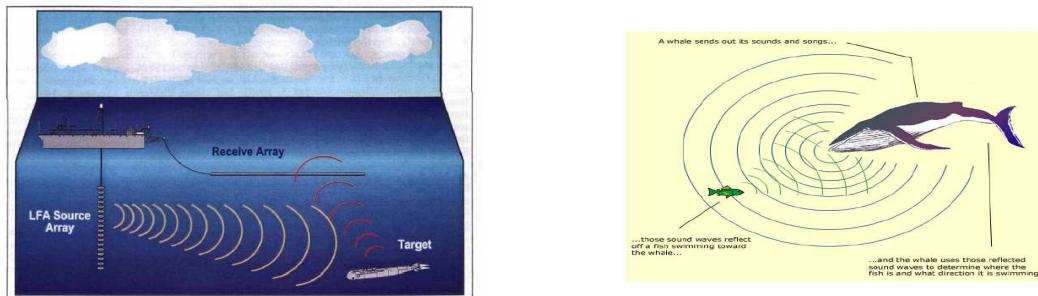


Fig.4.39 The figure depicts the working principle of SONAR (a) and in case of whales (b)

Sound Channels & Acoustic ducts in the ocean:

With reference to the typical vertical sound structure of the ocean in a layered medium we get The SOFAR channel, The Surface Duct, Shadow Zones which have a great practical significance for communication and detection in the under water.

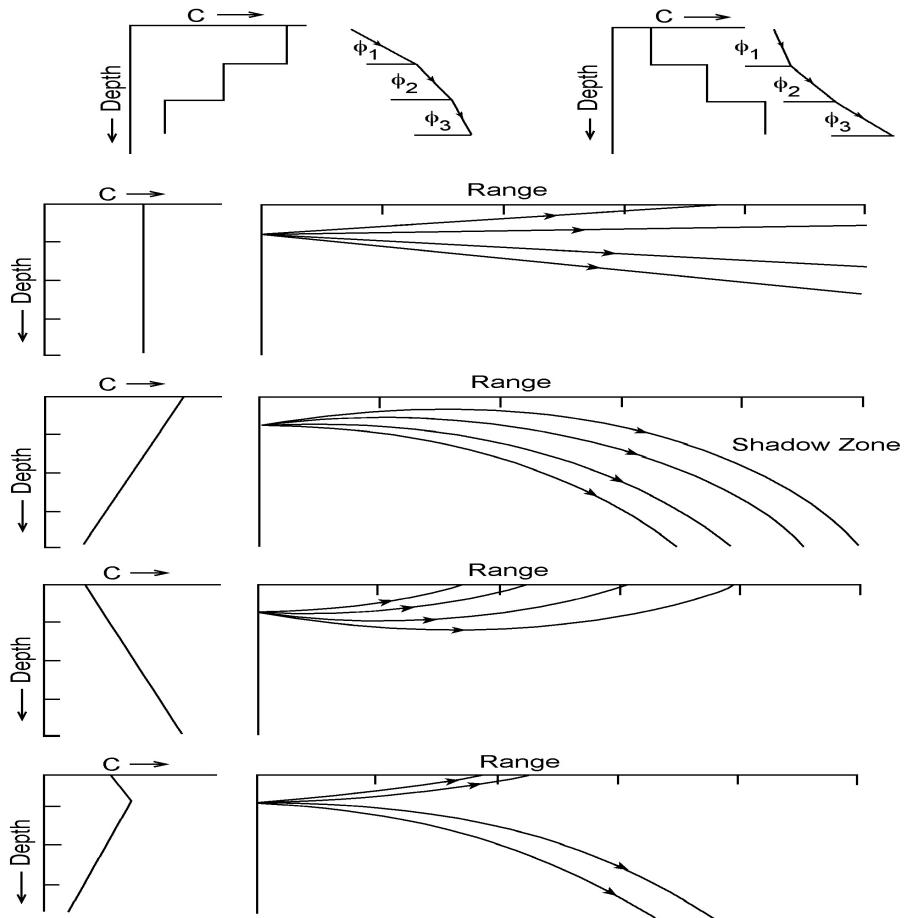


Fig.4.40. Ray paths with reference to the sound velocity

Figure 4.38 describes the distribution of ray paths with reference to the sound velocity and the resulting ducts and channels.

The combined effect of salinity, temperature and pressure produces a mid depth minimum in sound velocity in the oceans. The depth zone centered about this minimum is called the SOFAR channel. A sound generated in this zone escapes if it passes out at a large angle to the horizontal, however, if it strikes out from the source at low angles it is refracted back and forth along the center of the channel. Sound held captive in this manner is not dispersed spherically through the entire water body, but radiates outward from the source in relatively thin layer. Even small amounts of sound energy can traverse great horizontal distances when spherical divergence is prevented. In the SOFAR channel signals have been transmitted over distances as great as from Australia to Bermuda.

Sound refraction accounts for the development of shadow zones, sound channels and focusing and defocusing effects in underwater sound.

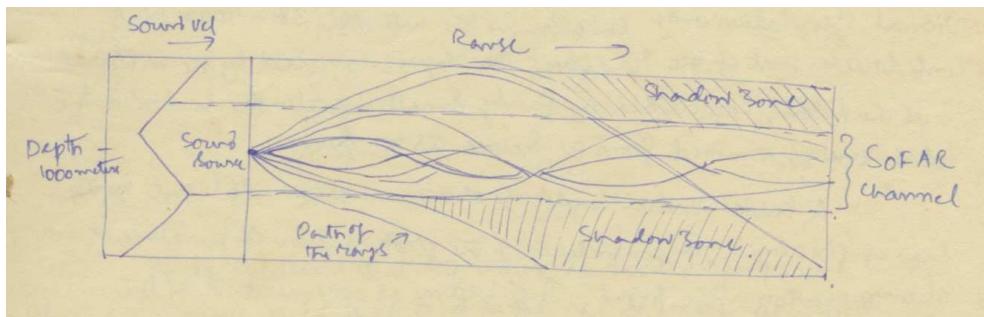


Fig.4. 41 Sound source placed at the inversion zone develops SOFAR CHANNEL

A shadow zone is produced when a positive velocity gradient lies above a negative velocity gradient and the sound source is in the positive gradient as in the above figures 4.32b & 4.39. The positive gradient may be due to an isothermal layer with the velocity of sound increasing with depth because of increasing pressure and in this case this sound ray will bend upward in the upper layers. Whereas below the isothermal layer the increase of sound velocity is not that quicker and as a result sound rays bend downward in the lower layer. The shadow zone is defined by the limiting ray which diverges at a particular depth corresponding to an abrupt change in the sound velocity profile.

According to ray theory, no sound energy enters the shadow zone. In actuality, the shadow zone is insonified at a greatly reduced intensity.

Two important sound channels exist in the oceans. The surface sound channel (Surface Acoustic duct) and the SOFAR channel as shown in the figures 4.36a and 4.36b respectively. Sound channels are caused by velocity minima in the velocity profile. Along the axis of the sound velocity, minimum rays are refracted upward and downward and are trapped and the sound is focused in the channels, and the attenuation in the channel is less than that predicted by the inverse square law as a result small charges exploded in the SOFAR channel (1500 meters) may be detected by hydrophones placed in the channel many thousands of kilometers away from the source.

In the case of figs 4.34b & 2nd and 4th of 4.40 above note that the sound energy does not reach deeper layers and so the sound is trapped in a duct near the surface and is known as the surface sound channel or surface duct.

As temperature decreases with depth, the sound rays are bent downwards and the sound energy is refracted into deeper layers. However, at depths around 1000 meters pressure increase dominates the temperature decrease. This leads to the development of SOFAR channel where the combined effects of downward refraction above mid depths and upward refraction below mid depths cause the sound energy to be trapped and concentrated in a channel. The sound can be transmitted to very long distances in SOFAR channel.

If we plot sound speed as a function of depth, we find that the speed usually has a minimum at a depth around 1000 m (figure 4.42 b). The depth at which minimum speed occurs is called the sound channel. It occurs in all oceans, and it usually reaches the surface at very high latitudes (Fig.4.42a). The sound channel is important because sound in the channel can travel

very far, sometimes half way around the earth. Here is how the channel works: Sound rays that begin to travel out of the channel are refracted back toward the center of the channel. Rays propagating upward at small angles to the horizontal are bent downward, and rays propagating downward at small angles to the horizontal are bent upward (figure 4.42a & b). Typical depths of the channel vary from 10 m to 1200 m depending on geographical area.

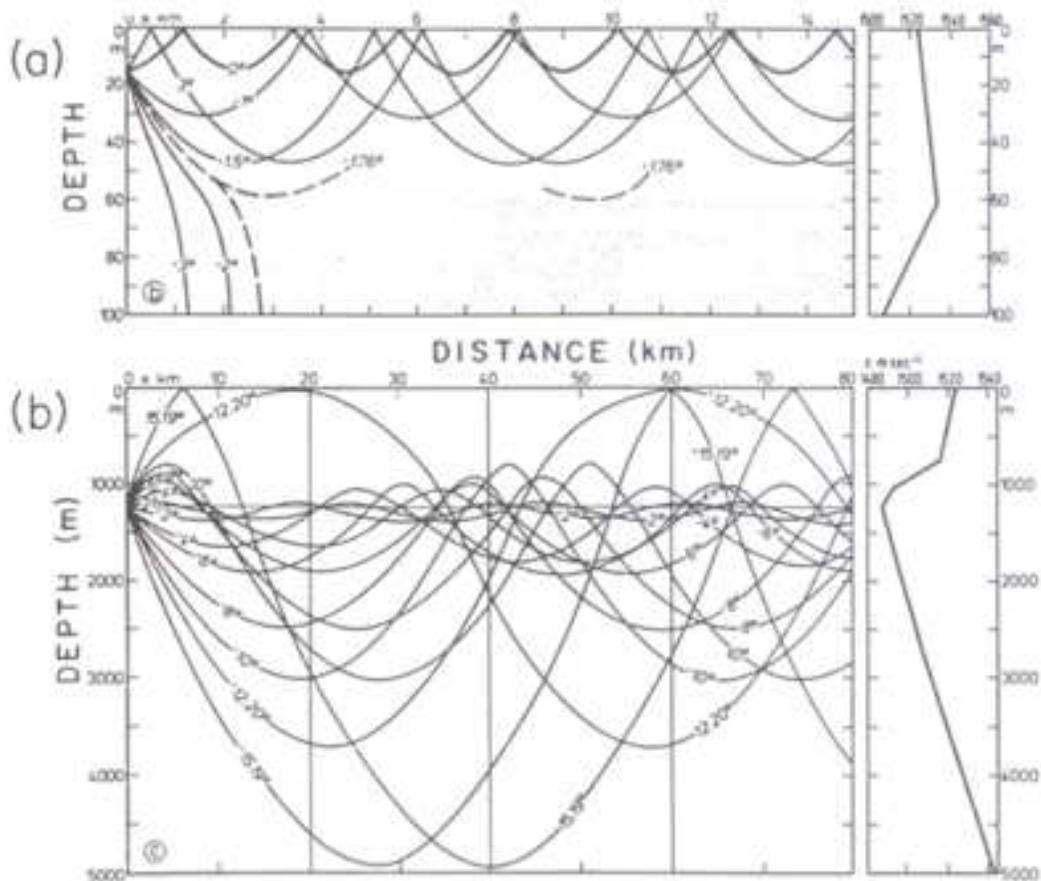


Fig.4. 42 a) Surface Sound channel b) SOFAR channel at higher latitudes

Very low frequency sounds in the sound channel, those with frequencies below 500 Hz have been detected at distances of megameters. In 1960, 15-Hz sound was sent from the explosions of 300 pounds (136 kg) of TNT off Perth, Western Australia (Fig. 4.43) at a depth of 1km near the sound axis. The acoustic signal was detected at Bermuda (32° N, 65° W), nearly halfway around the world, some 3.7 hours later which is the first demonstration of Global Acoustics. It appears that Bermuda is shadow zone. Then how it is detected at Bermuda? To answer this Munk suggests two paths one to South Africa and second to Bermuda (curves in fig.4.43).

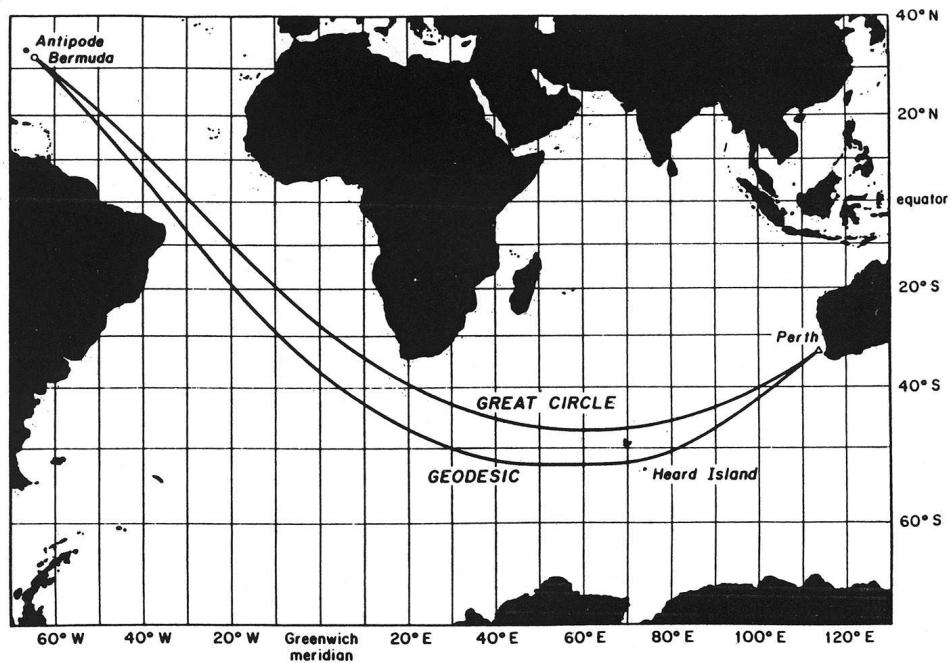


Fig. 4.43 Acoustic rays from Perth to Bermuda in 1960 experiment. Geodesic = allowing for earth's ellipticity. Great Circle = assumes spherical earth of mean radius. Dotted lines- constant sound speed. Bold lines-refracted paths (Munk et al 1994)

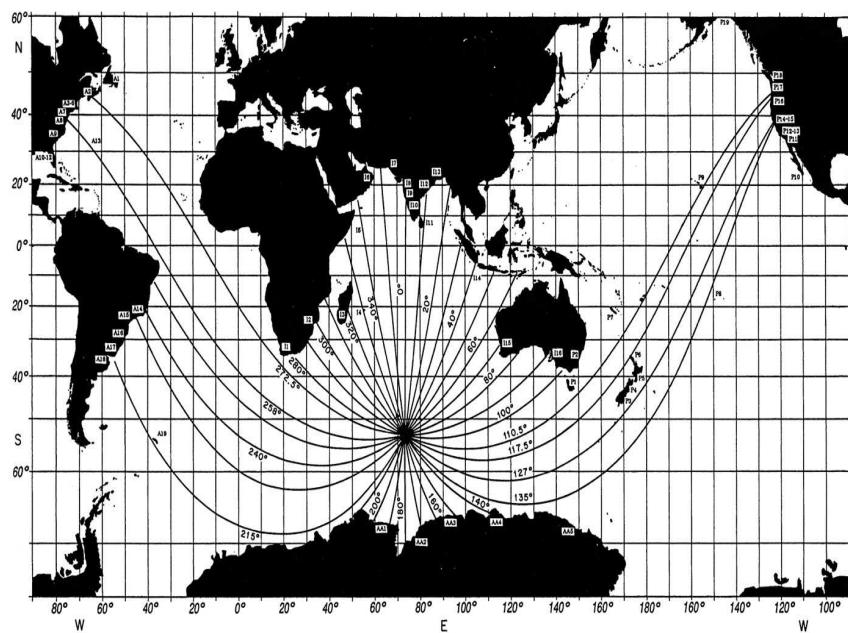


Fig.4. 44 Heard Island experiment in 1991(Munk et al 1994)

Later experiment showed that 57-Hz signals transmitted in the sound channel near Heard Island (75°E, 53°S) could be heard at Bermuda in the Atlantic and at Monterey, California in the Pacific (Fig.4.44). Heard Island experiment in 1991(Fig.4.44) thus demonstrated that signals could be detected on global scales using a coherent electrically induced sound source. Travel time to high accuracy (20 to 50 milliseconds) could be detected in N Atlantic and N Pacific. However, problems with sound scattering by the Antarctic Circumpolar Front making interpretation difficult

Munk suggested that since sound speed depended on temperature, the global warming may be detected by measuring change in the time delay between transmitter and receiver in the sound channel. Thus global warming was estimated in the sound channel (~ 1 km) at 30° latitude, to be about 5 milli degrees each year. This will cause a decrease in delay time by 0.1 to 0.2 seconds each year. Fig.4.39 shows the depth of minimum sound speed and its magnitude in the oceans in a general way.

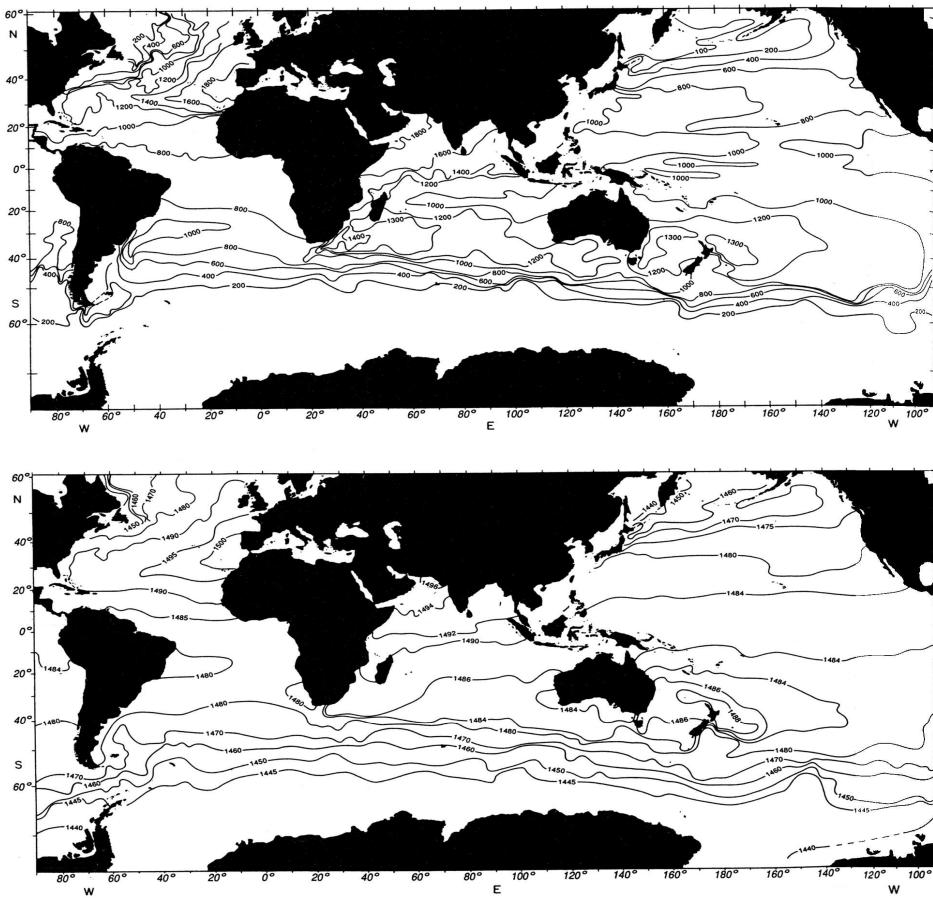


Fig.4. 39.Top: Depth (meters) of minimum sound speed. Bottom: Sound speed in m/s on the axial surface (above). The axis is typically depth of 1km, but outcrops at high northern and southern latitudes.

